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## RESEARCH ARTICLE

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### Key Points:

- A new process-based model incorporating microbial mechanisms was developed to quantify CH<sub>4</sub> emissions from natural wetlands
- The model was applied to 24 different wetlands globally to compare the simulated CH<sub>4</sub> emissions to observations
- The sensitivity analysis showed the significant impacts of parameters regulating DOC and acetate production, and acetoclastic methanogenesis on simulated CH<sub>4</sub> emissions

### Supporting Information:

- Supporting Information S1

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# A Microbial Functional Group-Based CH<sub>4</sub> Model Integrated Into a Terrestrial Ecosystem Model: Model Structure, Site-Level Evaluation, and Sensitivity Analysis

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**Abstract** Wetlands are one of the most important terrestrial ecosystems for land-atmosphere CH<sub>4</sub> exchange. A new process-based, biophysical model to quantify CH<sub>4</sub> emissions from natural wetlands was developed and integrated into a terrestrial ecosystem model (Integrated Biosphere Simulator). The new model represents a multisubstance system (CH<sub>4</sub>, O<sub>2</sub>, CO<sub>2</sub>, and H<sub>2</sub>) and describes CH<sub>4</sub> production, oxidation, and three transport processes (diffusion, plant-mediated transport, and ebullition). The new model uses several critical microbial mechanisms to represent the interaction of anaerobic fermenters and homoacetogens, hydrogenotrophic, and acetoclastic methanogens, and methanotrophs in CH<sub>4</sub> production and oxidation. We applied the model to 24 different wetlands globally to compare the simulated CH<sub>4</sub> emissions to observations and conducted a sensitivity analysis. The results indicated that (1) for most sites, the model was able to capture the magnitude and variation of observed CH<sub>4</sub> emissions under varying environmental conditions; (2) the parameters that regulate dissolved organic carbon and acetate production, and acetoclastic methanogenesis had the significant impact on simulated CH<sub>4</sub> emissions; (3) the representation of the process components of CH<sub>4</sub> cycling showed that CH<sub>4</sub> oxidation was about half or more of CH<sub>4</sub> production, and plant-mediated transport was the dominant pathway at most sites; and (4) the seasonality of simulated CH<sub>4</sub> emissions can be controlled by soil temperature, water table position, or combinations thereof.

**Plain Language Summary** CH<sub>4</sub> emission from wetlands is an important part of global carbon cycle. A new process-based model was developed to quantify the CH<sub>4</sub> emission from wetlands. The new model considered main microbial mechanisms and transport processes in wetland CH<sub>4</sub> cycling, and the modeled results matched the observed CH<sub>4</sub> emissions well at evaluation sites globally. A sensitivity analysis indicated the important role of parameters that controlled dissolved organic carbon and acetate production and acetoclastic methanogenesis. The assessment of process components of CH<sub>4</sub> cycling demonstrated the importance of CH<sub>4</sub> oxidation and plant-mediated transport in wetland CH<sub>4</sub> emission.

## 1. Introduction

CH<sub>4</sub> is well known to be an important greenhouse gas that has about 28 times the global warming potential of CO<sub>2</sub> over a 100-year scale (Myhre et al., 2013). The concentration of atmospheric CH<sub>4</sub> has increased from 722 ± 25 ppb in 1750 to 1,803 ± 2 ppb in 2011, about 2.5 times, since preindustrial periods (Hartmann et al., 2013). Natural wetlands, including wet soils, swamps, fens, bogs, and peatlands, are the largest

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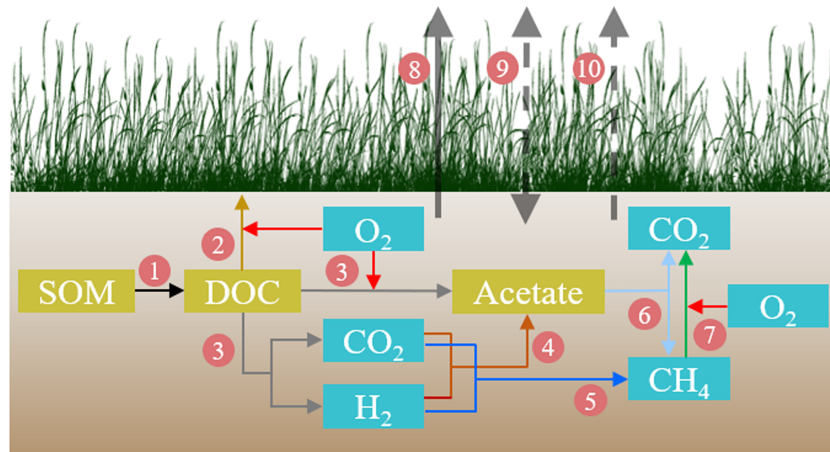
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individual natural CH<sub>4</sub> emissions source (Bridgman et al., 2013; Saunio et al., 2016). The estimation of CH<sub>4</sub> emissions from natural wetlands ranges from 153 to 227 Tg CH<sub>4</sub> per year for the 2003–2012 period, which accounts for 30% (top-down inversions) and 25% (bottom-up estimation) on the average of total CH<sub>4</sub> emissions (Saunio et al., 2016), meaning natural wetlands play an important role in the global CH<sub>4</sub> budget and climate change.

CH<sub>4</sub> emission from natural wetlands is a net balance between microbial CH<sub>4</sub> production and consumption. There exists two major CH<sub>4</sub> production mechanisms, hydrogenotrophic methanogenesis and acetoclastic methanogenesis (Conrad, 1999; Krüger et al., 2001); two CH<sub>4</sub> oxidation mechanisms, aerobic and anaerobic methanotrophy (Gerard & Chanton, 1993; Smemo & Yavitt, 2011); and three aggregated CH<sub>4</sub> transport pathways, plant-mediated transport, diffusion, and ebullition (Chanton, 2005; Mer & Roger, 2001; Whiting & Chanton, 1996), all of which are influenced by the availability and quality of substrate, soil temperature and pH, hydrological conditions, and vegetation composition, and so forth (Walter & Heimann, 2000). In anaerobic conditions, organic polymers are utilized by fermenters plus H<sub>2</sub>-producing acetogens to generate acetate, CO<sub>2</sub>, and H<sub>2</sub>, which become the substrates for hydrogenotrophic methanogens and acetoclastic methanogens and are ultimately transformed into CH<sub>4</sub>. After CH<sub>4</sub> is produced, it can be oxidized by methanotrophs to form CO<sub>2</sub> or be transported via vascular plants, molecular diffusion, and bubbles (Riley et al., 2011).

Numerous process-based models have been developed to quantify CH<sub>4</sub> emissions from natural wetlands during the past decades, including WMEM (Cao et al., 1996), the Arah model (Arah & Stephen, 1998), ecosys (Grant, 1998, 1999), the Walter model (Walter & Heimann, 2000), Wetland-DNDC (Zhang et al., 2002), the Kettunen model (Kettunen, 2003), PEATLAND-VU (Van Huissteden et al., 2006), TEM-CH<sub>4</sub> (Tang et al., 2010; Zhuang et al., 2004), DLEM (Tian et al., 2010), ORCHIDEE (Ringeval et al., 2010, 2011), LPJ-WHyMe (Wania et al., 2010), CLM4Me (Riley et al., 2011), VISIT (Ito & Inatomi, 2012), TRIPLEX-GHG (Zhu et al., 2014), CLM-Microbe (Xu et al., 2015), JSBACH-methane (Kaiser et al., 2017), TECO\_SPRUCE\_ME (Ma et al., 2017), HIMMELI (Raivonen et al., 2017), and the soil methane scheme of ISBA (Morel et al., 2019). These models explicitly consider CH<sub>4</sub>-related biogeochemical processes and their responses to environmental factors, and many of them simultaneously account for CH<sub>4</sub> production, oxidation, and three transport pathways. On the whole, these models share the following features: (1) the majority of models use a simple function directly associated with soil heterotrophic respiration, dissolved organic carbon (DOC), or environmental factors to estimate CH<sub>4</sub> production, lacking the representation of key microbial mechanisms; (2) the diffusion and ebullition processes are usually simulated as a threshold phenomenon whereby gas is released once the gas concentration exceeds a prescribed threshold concentration; and (3) many models have only been evaluated with a few in situ observations, without being compared against continuous CH<sub>4</sub> measurements collected over large spatial scales (only Riley et al., 2011 and Zhu et al., 2014 compared simulated and observed CH<sub>4</sub> emissions across different wetlands globally).

Using existing observations and studies, these models predict wetland CH<sub>4</sub> fluxes at ecosystem, regional and global scales. Yet recent studies have indicated that most previous models reproduced regional- to global-scale observations poorly (Bohn et al., 2015; Bohn & Lettenmaier, 2010). A recent model intercomparison project showed that the maximum estimate of global wetland CH<sub>4</sub> emissions is 264 Tg CH<sub>4</sub> per year, which is 1.89 times the minimum estimate of 141 Tg CH<sub>4</sub> per year (Melton et al., 2013). In addition, it is important to accurately quantify the response of global wetland CH<sub>4</sub> emissions to climate change, but Melton et al. (2013) found that the sign and magnitude of CH<sub>4</sub> emissions response to changes in temperature and precipitation vary among the models. Furthermore, only a few of the above models consider the interaction of different gases during the vertical migration of CH<sub>4</sub>; for example, the Arah model (Arah & Stephen, 1998), CLM4Me (Riley et al., 2011), and JSBACH-methane (Kaiser et al., 2017) account for CH<sub>4</sub> and O<sub>2</sub> simultaneously; LPJ-WHyMe (Wania et al., 2010), HIMMELI (Raivonen et al., 2017), and the soil methane scheme of ISBA (Morel et al., 2019) take CH<sub>4</sub>, O<sub>2</sub>, and CO<sub>2</sub> into consideration by keeping track of their dynamics throughout the process; and TEM-CH<sub>4</sub> (Tang et al., 2010) incorporates N<sub>2</sub>-related processes to make it a four-substance model that can resolve CH<sub>4</sub> biogeochemical cycling. Meanwhile, only three models incorporate the critical microbial mechanisms for CH<sub>4</sub> production and oxidation, including hydrogenotrophic methanogenesis, acetoclastic methanogenesis, and methanotrophy (e.g., Grant, 1998, 1999; Kettunen, 2003; Xu et al., 2015).



**Figure 1.** Schematic representation of key processes in the new wetland CH<sub>4</sub> model. The numbers in the red circle patterns are (1) soil organic matter decomposition, (2) aerobic respiration, (3) DOC decomposition (including fermentation and aerobic decomposition), (4) homoacetogenesis, (5) hydrogenotrophic methanogenesis, (6) acetoclastic methanogenesis, (7) CH<sub>4</sub> oxidation, (8) diffusion, (9) plant-mediated transport, and (10) ebullition. The red arrows represent the processes involving O<sub>2</sub>, including aerobic respiration, aerobic decomposition of DOC, and CH<sub>4</sub> oxidation.

In this study, we developed a new process-based model to quantify CH<sub>4</sub> emissions from natural wetlands, which considered the interaction of different gases, including CH<sub>4</sub>, O<sub>2</sub>, CO<sub>2</sub>, and H<sub>2</sub>, and incorporated the main microbial mechanisms related to CH<sub>4</sub> production and oxidation. Moreover, the new process-based model has been integrated in a terrestrial ecosystem model (Integrated Biosphere Simulator [IBIS]), which can reflect interactions between soil temperature, hydrology, vegetation, and CH<sub>4</sub> biogeochemical processes. The aims of this study are to (a) introduce a new wetland CH<sub>4</sub> emission model that includes the explicit description of the microbial mechanisms related to CH<sub>4</sub> generation, (b) show the capability of this model to simulate CH<sub>4</sub> emissions from natural wetlands by comparing with 24 site-level observations globally, and (c) assess the process components of wetland CH<sub>4</sub> cycling and the control on the seasonality of modeled CH<sub>4</sub> emissions.

## 2. Model Description

A new representation of natural wetland CH<sub>4</sub> emissions within a terrestrial ecosystem model, the IBIS, was developed for this study. The IBIS is a process-based, comprehensive model that describes land surface processes, terrestrial carbon balance, and vegetation dynamics. The main processes represented in the IBIS model include land surface physics, canopy physiology, phenology, vegetation structure and competition, and carbon and nitrogen cycling in the terrestrial biosphere and have been integrated into a single, physically consistent, with different time scales, modeling framework (Foley et al., 1996; Kucharik et al., 2000; Liu et al., 2014; Yuan et al., 2014); here, we only describe the development of novel wetland CH<sub>4</sub> dynamics. The new wetland CH<sub>4</sub> emission model (Figure 1) is a multidimensional substance system that simulates the biogeochemical processes related to CH<sub>4</sub>, O<sub>2</sub>, CO<sub>2</sub>, and H<sub>2</sub> and accounts for the transient, vertically resolved dynamics of these gases. The model introduces specific microbial mechanisms like anaerobic fermentation and homoacetogenesis, hydrogenotrophic methanogenesis, acetoclastic methanogenesis, and methanotrophy to account for CH<sub>4</sub> production and oxidation. To calculate net CH<sub>4</sub> emissions, the model also explicitly considers three different transport pathways including molecular diffusion, plant-mediated transport, and ebullition.

The transient reaction-diffusion equations governing the concentrations of CH<sub>4</sub>, O<sub>2</sub>, CO<sub>2</sub>, H<sub>2</sub>, and Ace (acetate) are:

$$\frac{\partial}{\partial t} C_{\text{CH}_4}(z, t) = \frac{\partial}{\partial z} F_{\text{diff, CH}_4} - Q_{\text{plant, CH}_4} - Q_{\text{ebull, CH}_4} + R_{\text{prod, CH}_4} - R_{\text{oxid, CH}_4}, \quad (1)$$

$$\frac{\partial}{\partial t} C_{\text{O}_2}(z, t) = \frac{\partial}{\partial z} F_{\text{diff, O}_2} - Q_{\text{plant, O}_2} - Q_{\text{ebull, O}_2} - R_{\text{aero}} - 2R_{\text{oxid, CH}_4}, \quad (2)$$



$$\frac{\partial}{\partial t} C_{\text{CO}_2}(z, t) = \frac{\partial}{\partial z} F_{\text{diff}, \text{CO}_2} - Q_{\text{plant}, \text{CO}_2} - Q_{\text{ebull}, \text{CO}_2} + R_{\text{prod}, \text{CO}_2} - R_{\text{cons}, \text{CO}_2}, \quad (3)$$

$$\frac{\partial}{\partial t} C_{\text{H}_2}(z, t) = \frac{\partial}{\partial z} F_{\text{diff}, \text{H}_2} - Q_{\text{plant}, \text{H}_2} - Q_{\text{ebull}, \text{H}_2} + R_{\text{prod}, \text{H}_2} - R_{\text{cons}, \text{H}_2}, \quad (4)$$

$$\frac{\partial}{\partial t} C_{\text{Ace}}(z, t) = R_{\text{prod}, \text{Ace}} - R_{\text{cons}, \text{Ace}}, \quad (5)$$

where  $C_X(z, t)$  is the concentration of compound  $X$  at soil layer depth  $z$  and time  $t$ ;  $F_{\text{diff}, X}$  is the diffusive flux of compound  $X$ ;  $Q_{\text{plant}, X}$  and  $Q_{\text{ebull}, X}$  are the transport rates of compound  $X$  via plant aerenchyma and ebullition, respectively;  $R_{\text{prod}, X}$  is the production rate of compound  $X$ ,  $R_{\text{oxid}, \text{CH}_4}$  is the  $\text{CH}_4$  oxidation rate;  $R_{\text{cons}, X}$  is the consumption rate of compound  $X$ ; and  $R_{\text{aero}}$  represents the aerobic respiration rate.

Water table depth is an important factor that separates the whole soil column into aerobic and anaerobic zones. In this study, site-level water table depth observations were used. We assume that  $\text{CH}_4$  is produced in the layers below the water table position and mainly consumed in the layers above the water table position (Walter & Heimann, 2000). The original IBIS included six soil layers with thicknesses of 0.1, 0.15, 0.25, 0.5, 1.0, and 2.0 m, with a total soil depth of 4 m (Foley et al., 1996). In order to represent water table depth specifically, we divided the soil column into nine layers, with the first five layers having a thickness of 0.1 m and the other four layers having thicknesses of 0.2, 0.3, 0.5, and 0.5 m, respectively. We only considered the  $\text{CH}_4$ -related biogeochemical processes occurring in the nine soil layers. The detailed algorithms are described as follows.

## 2.1. $\text{CH}_4$ Production

### 2.1.1. Available Carbon Decomposition

Methanogens prefer fresh organic matter as their substrate (Joabsson & Christensen, 2001), so the initial carbon source available for methanogens mainly comprises easily decomposed plant litterfall and root exudates. Given that DOC is a key intermediate for anaerobic fermentation and is converted from hydrolysis products and eventually decomposed into acetate,  $\text{CO}_2$ , and  $\text{H}_2$ , we define it to be the original, available carbon for methanogens and use a simple formula for its calculation:

$$\text{DOC} = K_{\text{cpool}} \times \frac{\text{cpool}}{dz} \times f_T(\text{DOCprodQ}_{10}) \times f_{\text{moist}}. \quad (6)$$

Here, DOC is the dissolved organic carbon concentration ( $\text{mol m}^{-3}$ ),  $K_{\text{cpool}}$  represents the ratio of DOC to soil organic carbon, cpool is the soil organic carbon content ( $\text{mol m}^{-2}$ ),  $dz$  is the layer thickness (m),  $\text{DOCprodQ}_{10}$  is the temperature sensitivity of DOC production,  $f_T$  and  $f_{\text{moist}}$  are soil temperature and moisture factors, respectively. We use the approach adopted by Wania et al. (2010) and Raivonen et al. (2017) to distribute the available carbon for methanogens to all soil layers according to the root fraction  $f_{\text{root}}$ , which is calculated in the IBIS.

Under anaerobic conditions, the available carbon is fermented into acetate,  $\text{CO}_2$ , and  $\text{H}_2$ , which is governed by the following equation:

$$\text{DOC} \rightarrow 0.67\text{Ace} + 0.33\text{CO}_2 + 0.11\text{H}_2. \quad (7)$$

Thus, the fermentation is presented as:

$$\text{DOCprodAce} = V_{\text{DOCprodAce}, \text{max}} \times \frac{\text{DOC}}{K_{\text{DOCprodAce}} + \text{DOC}} \times f_T(\text{AceprodQ}_{10}) \times f_{\text{pH}}, \quad (8)$$

$$\text{DOCprodCO}_2 = 0.5 \times \text{DOCprodAce}, \quad (9)$$

$$\text{DOCprodH}_2 = \frac{1}{6} \times \text{DOCprodAce}. \quad (10)$$

Under aerobic conditions, the available carbon is decomposed into acetate and  $\text{CO}_2$ , which can be calculated as

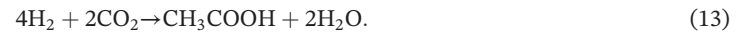
$$\text{DOCprodAce} = V_{\text{DOCprodAce,max}} \times \frac{\text{DOC}}{K_{\text{DOCprodAce}} + \text{DOC}} \times \frac{[\text{O}_2]}{K_{\text{AceprodO}_2} + [\text{O}_2]} \times f_T(\text{AceprodQ}_{10}) \times f_{\text{pH}}, \quad (11)$$

$$\text{DOCprodCO}_2 = 0.5 \times \text{DOCprodAce}. \quad (12)$$

Here,  $\text{DOCprodAce}$ ,  $\text{DOCprodCO}_2$ , and  $\text{DOCprodH}_2$  are the production rate ( $\text{mol m}^{-3} \text{d}^{-1}$ ) of acetate,  $\text{CO}_2$ , and  $\text{H}_2$ , respectively.  $V_{\text{DOCprodAce,max}}$  is the maximum acetate production rate ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $[\text{O}_2]$  is the  $\text{O}_2$  concentration ( $\text{mol m}^{-3}$ ), and  $K_{\text{DOCprodAce}}$  and  $K_{\text{AceprodO}_2}$  are the half saturation coefficients ( $\text{mol m}^{-3}$ ) for DOC and  $\text{O}_2$ , respectively.  $\text{AceprodQ}_{10}$  is the temperature sensitivity of acetate production, and  $f_{\text{pH}}$  is the soil pH factor.

### 2.1.2. Homoacetogenesis and Hydrogenotrophic Methanogenesis

The products  $\text{CO}_2$  and  $\text{H}_2$  are substrates for homoacetogens and hydrogenotrophic methanogens (Grant, 1998; Grant & Roulet, 2002). At low temperature,  $\text{CO}_2$  and  $\text{H}_2$  appear to be the main substrates for homoacetogens, which use  $\text{CO}_2$  as the electron acceptor to convert  $\text{H}_2$  to acetate (chemolithotrophic acetogenesis/homoacetogenesis) (Kotsyurbenko et al., 2001; Liu & Conrad, 2011; Schulz & Conrad, 1996). This process is governed by the following equation:



Thus, acetate production from homoacetogenesis is calculated as

$$\text{H}_2\text{prodAce} = V_{\text{H}_2\text{prodAce,max}} \times \text{Homoacetogens} \times \frac{[\text{H}_2]}{K_{\text{H}_2\text{prodAce}} + [\text{H}_2]} \times \frac{[\text{CO}_2]}{K_{\text{CO}_2\text{prodAce}} + [\text{CO}_2]} \times f_{T1} \times f_{\text{pH}}. \quad (14)$$

Here,  $\text{H}_2\text{prodAce}$  is the acetate production rate from homoacetogenesis ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $V_{\text{H}_2\text{prodAce,max}}$  is the maximum acetate production rate ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $\text{Homoacetogens}$  is the microbial biomass of homoacetogens ( $\text{mol m}^{-3}$ ),  $[\text{H}_2]$  and  $[\text{CO}_2]$  are the concentrations of  $\text{H}_2$  and  $\text{CO}_2$  ( $\text{mol m}^{-3}$ ),  $K_{\text{H}_2\text{prodAce}}$  and  $K_{\text{CO}_2\text{prodAce}}$  are the half saturation coefficients ( $\text{mol m}^{-3}$ ) for  $\text{H}_2$  and  $\text{CO}_2$ , respectively, and  $f_{T1}$  represents the soil temperature factor.

In contrast, at high temperature, produced  $\text{CO}_2$  and  $\text{H}_2$  have usually been found to be suitable substrates for hydrogenotrophic methanogens (Hattroi, 2008), which use  $\text{CO}_2$  and  $\text{H}_2$  to generate  $\text{CH}_4$  (hydrogenotrophic methanogenesis). This process is governed by the following equation:



Thus,  $\text{CH}_4$  production from hydrogenotrophic methanogenesis is calculated as

$$\text{H}_2\text{prodCH}_4 = V_{\text{H}_2\text{prodCH}_4,\text{max}} \times \text{H}_2\text{methanogens} \times \frac{[\text{H}_2]}{K_{\text{H}_2\text{prodCH}_4} + [\text{H}_2]} \times \frac{[\text{CO}_2]}{K_{\text{CO}_2\text{prodCH}_4} + [\text{CO}_2]} \times f_{T2} \times f_{\text{pH}}. \quad (16)$$

Here,  $\text{H}_2\text{prodCH}_4$  is the  $\text{CH}_4$  production rate from hydrogenotrophic methanogenesis ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $V_{\text{H}_2\text{prodCH}_4,\text{max}}$  is the maximum  $\text{CH}_4$  production rate ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $\text{H}_2\text{methanogens}$  is the microbial biomass of hydrogenotrophic methanogens ( $\text{mol m}^{-3}$ ),  $K_{\text{H}_2\text{prodCH}_4}$  and  $K_{\text{CO}_2\text{prodCH}_4}$  are the half saturation coefficients ( $\text{mol m}^{-3}$ ) for  $\text{H}_2$  and  $\text{CO}_2$ , respectively, and  $f_{T2}$  represents the soil temperature factor.

### 2.1.3. Acetoclastic Methanogenesis

Product acetate is the substrate for acetoclastic methanogens (Grant, 1998; Grant & Roulet, 2002).  $\text{CH}_4$  production from acetoclastic methanogenesis is governed by the following equation:



Thus,  $\text{CH}_4$  production is calculated as

$$\text{AceprodCH}_4 = K_{\text{CH}_4\text{prod}} \times (1 - \text{Grow}_{\text{Acemethanogens}}) \times \text{Acecons}. \quad (18)$$

The acetate consumption is calculated as

$$\text{Acecons} = V_{\text{Acecons,max}} \times \text{Acemethanogens} \times \frac{\text{Ace}}{K_{\text{AceprodCH}_4} + \text{Ace}} \times f_T(\text{CH}_4\text{prod}Q_{10}) \times f_{\text{pH}}. \quad (19)$$

Here,  $\text{AceprodCH}_4$  is the  $\text{CH}_4$  production rate from acetate consumption ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $K_{\text{CH}_4\text{prod}}$  is the  $\text{CH}_4$  production ratio,  $\text{Grow}_{\text{Acemethanogens}}$  is the growth efficiency of acetoclastic methanogens,  $\text{Acecons}$  is the acetate consumption rate for respiration by acetoclastic methanogens ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $V_{\text{Acecons,max}}$  is the maximum acetate consumption rate ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $\text{Acemethanogens}$  is the microbial biomass of acetoclastic methanogens ( $\text{mol m}^{-3}$ ),  $\text{Ace}$  is the acetate concentration ( $\text{mol m}^{-3}$ ),  $K_{\text{AceprodCH}_4}$  is the half saturation coefficient ( $\text{mol m}^{-3}$ ) for acetate, and  $\text{CH}_4\text{prod}Q_{10}$  is the temperature sensitivity of  $\text{CH}_4$  production during acetate consumption.

## 2.2. $\text{CH}_4$ Oxidation

The  $\text{CH}_4$  produced by hydrogenotrophic methanogens and acetoclastic methanogens is the substrate for methanotrophs (Grant, 1999; Grant & Roulet, 2002). In this process, heterotrophic methanotrophs are the main microbial functional group (Riley et al., 2011). We use double Michaelis-Menten kinetics to represent  $\text{CH}_4$  oxidation, which is governed by the following equation:



Thus, the  $R_{\text{oxid,CH}_4}$  ( $\text{mol m}^{-3} \text{d}^{-1}$ ) is calculated as

$$R_{\text{oxid,CH}_4} = V_{\text{CH}_4\text{oxid,max}} \times \text{Methanotrophs} \times \frac{[\text{CH}_4]}{K_{\text{CH}_4\text{oxidCH}_4} + [\text{CH}_4]} \times \frac{[\text{O}_2]}{K_{\text{CH}_4\text{oxidO}_2} + [\text{O}_2]} \times f_T(\text{CH}_4\text{oxid}Q_{10}) \times f_{\text{pH}}. \quad (21)$$

Here,  $V_{\text{CH}_4\text{oxid,max}}$  is the maximum  $\text{CH}_4$  oxidation rate ( $\text{mol m}^{-3} \text{d}^{-1}$ ),  $\text{Methanotrophs}$  is the microbial biomass of methanotrophs ( $\text{mol m}^{-3}$ ),  $[\text{CH}_4]$  is the  $\text{CH}_4$  concentration ( $\text{mol m}^{-3}$ ),  $K_{\text{CH}_4\text{oxidCH}_4}$  and  $K_{\text{CH}_4\text{oxidO}_2}$  are the half saturation coefficients ( $\text{mol m}^{-3}$ ) for  $\text{CH}_4$  and  $\text{O}_2$ , respectively, and  $\text{CH}_4\text{oxid}Q_{10}$  is the temperature sensitivity of  $\text{CH}_4$  oxidation.

$\text{O}_2$  is not only used to oxidize  $\text{CH}_4$  but also consumed by aerobic respiration. We assume that part of the available carbon will be used by aerobic microbes, and this process requires  $\text{O}_2$ , so refer to Kettunen (2003); the  $R_{\text{aero}}$  ( $\text{mol m}^{-3} \text{d}^{-1}$ ) is calculated as

$$R_{\text{aero}} = K_{\text{aer}} \times \frac{\text{DOC}}{K_{\text{aerDOC}} + \text{DOC}} \times \frac{[\text{O}_2]}{K_{\text{aerO}_2} + [\text{O}_2]} \times f_T(\text{DOCprod}Q_{10}) \times f_{\text{pH}}. \quad (22)$$

Here,  $K_{\text{aer}}$  is the  $\text{O}_2$  consumption rate by aerobic respiration ( $\text{mol m}^{-3} \text{d}^{-1}$ ), and  $K_{\text{aerDOC}}$  and  $K_{\text{aerO}_2}$  are the half saturation coefficients ( $\text{mol m}^{-3}$ ) for DOC and  $\text{O}_2$ , respectively.

In the above processes,  $R_{\text{prod,X}}$  ( $\text{mol m}^{-3} \text{d}^{-1}$ ) is the sum of produced compound X, and  $R_{\text{cons,X}}$  ( $\text{mol m}^{-3} \text{d}^{-1}$ ) is the sum of consumed compound X.

## 2.3. $\text{CH}_4$ Transportation

### 2.3.1. Diffusion

Molecular diffusive flux  $F_{\text{diff,X}}$  ( $\text{mol m}^{-2} \text{d}^{-1}$ ) within the soil profile depends on the vertical concentration gradient and the diffusion coefficients of compound X. Fick's first law is applied to calculate the diffusive flux (Walter & Heimann, 2000).

$$F_{\text{diff,X}} = D_X \frac{\partial C_X}{\partial z}. \quad (23)$$

Here,  $D_X$  is the effective diffusivity of compound X ( $\text{m}^2 \text{s}^{-1}$ ). We also use the constant reduction factors  $f_{\text{D,w}}$  and  $f_{\text{D,a}}$  to calculate the effective diffusivities in water or air (Raivonen et al., 2017), respectively.

$$D_{X,w} = f_{D,w} \times D_X^{\text{water}}, \quad (24)$$

$$D_{X,a} = f_{D,a} \times D_X^{\text{air}}. \quad (25)$$

Here,  $D_{X,w}$  and  $D_{X,a}$  represent the effective diffusivities of compound X in water and air ( $\text{m}^2 \text{s}^{-1}$ ), and  $D_X^{\text{water}}$  and  $D_X^{\text{air}}$  are the free-water and free-air diffusivities ( $\text{m}^2 \text{s}^{-1}$ ), respectively.

At the air-water interface, diffusivities can vary by at least four orders of magnitude. The method adopted by Wania et al. (2010) is used to calculate gas fluxes from the top soil layer into the atmosphere.

$$F_{\text{diff},X} = -\varphi_X \times (C_{\text{surf},X} - C_{\text{eq},X}). \quad (26)$$

Here,  $C_{\text{surf},X}$  is the concentration of compound X in the top soil layer ( $\text{mol m}^{-3}$ ), and  $C_{\text{eq},X}$  is the equilibrium concentration of compound X in the atmosphere ( $\text{mol m}^{-3}$ ).  $\varphi_X$  represents the transfer velocity of compound X ( $\text{cm hr}^{-1}$ ). We use a normalized transfer velocity  $\varphi_{600}$  (Cole & Caraco, 1998) to calculate the  $\varphi_X$ .

$$\varphi_X = \varphi_{600} \times \left( \frac{Sc_X}{600} \right)^n. \quad (27)$$

The  $\varphi_{600}$  ( $\text{cm hr}^{-1}$ ) is calculated as

$$\varphi_{600} = 2.07 + 0.215 \times U_{10}^{1.7}. \quad (28)$$

Here,  $Sc_X$  is the Schmidt number of compound X, 600 is the Schmidt number for  $\text{CO}_2$  at  $20^\circ\text{C}$ ,  $n = -0.5$ , and  $U_{10}$  is the wind speed at 10-m height ( $\text{m s}^{-1}$ ). In this study, we also assume that wind speed can be ignored within the wetland vegetation so use a constant value of 0 for  $U_{10}$ . The unit of  $\varphi_X$  is centimeter per hour, which can be further transformed into meter per day.

For the concentration  $C_{\text{eq},X}$ , which is in equilibrium with the gas partial pressure  $PP_X$  (Pa) and can be computed as:

$$C_{\text{eq},X} = PP_X \times H_X. \quad (29)$$

Here,  $H_X$  is the Henry's law constant for compound X ( $\text{mol m}^{-3} \text{Pa}^{-1}$ ). Based on the above equations, the diffusive flux for compound X can be obtained. The solution for diffusion within the soil column is obtained using the Crank-Nicholson scheme (Press et al., 1996).

### 2.3.2. Plant Transport

Many wetland vascular plants develop aerenchyma in response to the inundation environment. These tissues can act as conduits for the transport of  $\text{CH}_4$ ,  $\text{O}_2$ ,  $\text{CO}_2$ , and  $\text{H}_2$  between the soil and atmosphere. Thus, plant-mediated transport is a diffusion process through the aerenchyma and driven by the specific gas concentration gradient (Riley et al., 2011). We refer to the approach adopted by Stephen et al. (1998) to calculate the plant transport rate  $Q_{\text{plant},X}$  ( $\text{mol m}^{-3} \text{d}^{-1}$ ).

$$Q_{\text{plant},X} = \frac{D_{\text{air},X}}{\tau} \times \varepsilon(z) \times \frac{C_X(z,t) - C_{\text{eq},X}}{z}. \quad (30)$$

Here,  $\varepsilon(z)$  is the density of cross-sectional area of root endings at depth  $z$  ( $\text{m}^2 \text{m}^{-3}$ ), and  $\tau$  is the root tortuosity. The effective diffusivities in air are used as the diffusion coefficients inside roots for each gas. Similar to Raivonen et al. (2017),  $\varepsilon(z)$  is formulated as

$$\varepsilon(z) = a_{\text{mA}} \times \frac{f_{\text{root}}}{dz} \times \frac{\text{LAI}}{\text{SLA}}. \quad (31)$$

Here,  $a_{\text{mA}}$  is the cross-sectional area of root endings per root biomass ( $\text{m}^2 \text{kg}^{-1}$ ), SLA represents the specific leaf area ( $\text{m}^2 \text{kg}^{-1}$ ), and LAI is the leaf area index ( $\text{m}^2 \text{m}^{-2}$ ), which is modeled in IBIS.

### 2.3.3. Ebullition

Ebullition is a relatively rapid process and only occurs in water-filled soil when the total partial pressure of dissolved gases exceeds the sum of atmospheric and hydrostatic pressure. Our implementation of ebullition follows that of Tang et al. (2010) and Raivonen et al. (2017), which is a new algorithm based on hydrostatic equilibrium rather than concentration threshold. The ebullition algorithm considers the concentrations of CH<sub>4</sub>, O<sub>2</sub>, CO<sub>2</sub>, H<sub>2</sub>, and N<sub>2</sub>, when the criterion for bubble formation is reached; such that when

$$\sum_X PP_X(z) > P_{\text{atm}} + P_{\text{hyd}}, \quad (32)$$

ebullition occurs. Here,  $P_{\text{atm}}$  and  $P_{\text{hyd}}$  are the atmospheric and hydrostatic pressure (Pa), respectively.

The fraction of ebullition  $f_{\text{ebull}}(z)$  is calculated as

$$f_{\text{ebull}}(z) = \frac{\sum_X PP_X(z) - (P_{\text{atm}} + P_{\text{hyd}})}{\sum_X PP_X(z)}. \quad (33)$$

We also use the ebullition rate constant  $k$  (d<sup>-1</sup>) in the equation, so the ebullition rate  $Q_{\text{ebull},X}$  (mol m<sup>-3</sup> d<sup>-1</sup>) of compound  $X$  is calculated as

$$Q_{\text{ebull},X} = k \times \frac{\sigma \times f_{\text{ebull}}(z) \times PP_X(z)}{RT}. \quad (34)$$

Here,  $\sigma$  is the porosity,  $R$  is the universal gas constant (J mol<sup>-1</sup> K<sup>-1</sup>) and  $T$  is the soil temperature (K).

The total ebullition flux released into either the atmosphere or soil is determined by water table depth. If the position of the water table is below the soil surface, the bubbles of gas are transported into the overlying air-filled soil layer and are subsequently diffused into the soil or plant aerenchyma. Otherwise, they are directly released into the atmosphere.

The detailed descriptions and baseline values for all parameters of the CH<sub>4</sub> model are listed in Table 1. The algorithms used to describe the dynamics of four microbial groups, the coefficients associated with gas transport processes (including diffusion coefficients, Henry law constants, and Schmidt numbers), and the specific expressions for environmental factors are presented in Appendices A, B, and C, respectively.

## 3. Data and Methods

### 3.1. Observations

Continuous observations of CH<sub>4</sub> emissions were compiled from 24 natural wetland sites, covering tropical, temperate, and boreal regions. Detailed information regarding these sites has been reported in previous studies or field work, so site information is briefly introduced in Tables 2–4, including location, wetland types, dominant vegetation, years of observation, and the measurement methods.

### 3.2. Model Forcing Data

The forcing data for this model mainly include daily climate data, daily water table depth, and soil carbon data. We used the Modern-Era Retrospective Analysis for Research and Applications data set (Gelaro et al., 2017) and the Climate Prediction Center Global Unified Precipitation data provided by the National Oceanic and Atmospheric Administration/Oceanic and Atmospheric Research/Earth System Research Laboratory Physical Sciences Division, Boulder, Colorado, United States (<https://www.esrl.noaa.gov/psd/>) to drive the model. The Modern-Era Retrospective Analysis for Research and Applications data provided the daily maximum, minimum, and average air temperature, relative humidity, and wind speed at 2 m, and the Climate Prediction Center Global Unified Precipitation data provided the daily precipitation. The soil carbon data for model initialization were obtained from the Global Gridded Surfaces of Selected Soil Characteristics (International Geosphere-Biosphere Programme Data and Information System) data set (Global Soil Data Task Group, 2000), and the soil pH was obtained from the soil properties data set of the Digital Soil Map of the World by Land and Water Development Division, Food and Agriculture Organization (<http://www.fao.org/geonetwork/srv/en/metadata.show?id=14116>). A spin-up of 200 years



**Table 1**  
Major Parameters in CH<sub>4</sub>-Related Processes

Index	Parameter	Value	Range	Units	Descriptions	References
P1	K <sub>cpool</sub>	0.02	0.0001–0.02		Ratio of dissolved organic carbon to soil organic carbon	
P2	V <sub>DOCprodAce,max</sub>	0.5	0.3–0.7	mol m <sup>-3</sup> d <sup>-1</sup>	Maximum acetate production rate from fermentation	Grant (1998)
P3	K <sub>DOCprodAce</sub>	10.0	5–15	mol m <sup>-3</sup>	Half saturation coefficient	
P4	K <sub>AceprodO<sub>2</sub></sub>	0.04	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	Kettunen (2003)
P5	V <sub>H<sub>2</sub>prodAce,max</sub>	0.15	0.01–0.3	mol m <sup>-3</sup> d <sup>-1</sup>	Maximum acetate production rate from homoacetogenesis	
P6	V <sub>H<sub>2</sub>prodCH<sub>4</sub>,max</sub>	0.15	0.01–0.3	mol m <sup>-3</sup> d <sup>-1</sup>	Maximum CH <sub>4</sub> production rate from hydrogenotrophic methanogenesis	Grant (1998)
P7	V <sub>Acecons,max</sub>	0.5	0.3–0.7	mol m <sup>-3</sup> d <sup>-1</sup>	Maximum acetate consumption rate by acetoclastic methanogenesis	Grant (1998)
P8	V <sub>CH<sub>4</sub>oxid,max</sub>	0.5	0.3–0.7	mol m <sup>-3</sup> d <sup>-1</sup>	Maximum CH <sub>4</sub> oxidation rate	Grant (1999)
P9	K <sub>H<sub>2</sub>prodAce</sub>	0.01	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	
P10	K <sub>CO<sub>2</sub>prodAce</sub>	0.05	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	
P11	K <sub>H<sub>2</sub>prodCH<sub>4</sub></sub>	0.01	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	
P12	K <sub>CO<sub>2</sub>prodCH<sub>4</sub></sub>	0.05	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	
P13	K <sub>AceprodCH<sub>4</sub></sub>	0.05	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	Kettunen (2003)
P14	K <sub>CH<sub>4</sub>prod</sub>	0.5	0.3–0.7	mol mol <sup>-1</sup>	CH <sub>4</sub> production ratio	Kettunen (2003)
P15	K <sub>CH<sub>4</sub>oxidCH<sub>4</sub></sub>	0.05	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	Kettunen (2003)
P16	K <sub>CH<sub>4</sub>oxidO<sub>2</sub></sub>	0.02	0.01–0.1	mol m <sup>-3</sup>	Half saturation coefficient	Kettunen (2003)
P17	f <sub>D,w</sub>	0.8	0.7–0.9		Reduction factor for diffusion in water-filled peat	Raivonen et al. (2017)
P18	f <sub>D,a</sub>	0.8	0.7–0.9		Reduction factor for diffusion in air-filled peat	Raivonen et al. (2017)
P19	τ	1.5	1–2		Root tortuosity	Stephen et al. (1998)
P20	a <sub>mA</sub>	0.085	0.01–0.1	m <sup>2</sup> kg <sup>-1</sup>	Root ending area per root dry biomass	Stephen et al. (1998)
P21	SLA	20	15–25	m <sup>2</sup> kg <sup>-1</sup>	Specific leaf area	Raivonen et al. (2017)
P22	σ	0.8	0.7–0.9		Peat porosity	Raivonen et al. (2017)
P23	k	0.0001	0.0001–0.0003	d <sup>-1</sup>	Time constant of ebullition	
P24	K <sub>aer</sub>	0.1	0.1–0.2	mol m <sup>-3</sup> d <sup>-1</sup>	O <sub>2</sub> consume rate from aerobic respiration	
P25	K <sub>aerDOC</sub>	10.0	5–15	mol m <sup>-3</sup>	Half saturation coefficient	
P26	K <sub>aerO<sub>2</sub></sub>	0.22	0.01–0.3	mol m <sup>-3</sup>	Half saturation coefficient	
P27	DOCprodQ <sub>10</sub>	2.5	1–5		Temperature sensitivity of DOC production	Kettunen (2003)
P28	AceprodQ <sub>10</sub>	2.5	1–5		Temperature sensitivity of acetate production	Kettunen (2003)
P29	CH <sub>4</sub> prodQ <sub>10</sub>	2.5	1–5		Temperature sensitivity of CH <sub>4</sub> production	Kettunen (2003)
P30	CH <sub>4</sub> oxidQ <sub>10</sub>	2.5	1–5		Temperature sensitivity of CH <sub>4</sub> oxidation	Kettunen (2003)

(repeat the climate data of observed years) was done to obtain the modeled soil temperature and moisture, and the last cycle of observed years was used to simulate the CH<sub>4</sub> emissions. The final output of the model was the daily fluxes of gases between soil and atmosphere.

### 3.3. Sensitivity Analysis

To characterize the sensitivity of the modeled CH<sub>4</sub> emissions to input parameters (Table 1), the polynomial chaos expansion (PCE)-based Sobol sensitivity indices method was employed to assess the response of model output to changes in parameters. The Sobol indices (Sobol, 1993) are convenient representations of the model sensitivity to its parameters and correspond to the variance-based decomposition that aims to decompose the total variance of the model into the sum of the variances of each input variable or their interactions (Ricciuto et al., 2018; Sudret, 2008; Wang et al., 2016). The PCE is a powerful probabilistic technique that uses the orthogonal stochastic polynomials in the random inputs to provide a functional approximation of the model output (Crestaux et al., 2009; Marelli et al., 2019). The Sobol indices are traditionally computed by Monte Carlo simulation, which makes them difficult to apply with computationally expensive models (Sudret, 2008). Instead, the PCE-based Sobol indices method requires less computational effort and is more applicable for the computationally expensive models (Marelli et al., 2019) and has been applied to some sensitivity studies of parameters in earth system model or land surface model (Ricciuto et al., 2018; Shi et al., 2019). The PCE-based Sobol indices method is briefly introduced as follows:

For an input random vector  $X = \{X_1, \dots, X_k\}$ , the PCE of model output  $Y = M(X)$  can be established as

**Table 2**  
Description of Sites in Boreal Regions

No.	Site name	Location	Wetland type	Dominant vegetation	Time	Method	References
1	Zackenbergl Greenland	74°30'N, 21°00'W	Arctic Fen	<i>Eriophorum scheuchzeri</i> , <i>Carex stans</i> , and <i>Duportia psilosantha</i>	2006–2010	Automatic chamber	Mastepanov et al. (2013)
2	Siikaneva, Finland	61°50'N, 24°12'E	Boreal minerotrophic fen	Sedges, Rannoch rush and Mosses	2005–2015	Eddy covariance	Rinne et al. (2018)
3	Lompolojätkä, Finland	67°59.8'N, 24°12.5'E	Open, nutrient-rich sedge fen	<i>Betula nana</i> , <i>Menyanthes trifoliata</i> , <i>Salix lapponum</i> , and <i>Carex ssp.</i>	2006–2010	Eddy covariance	Raivonen et al. (2017)
4	Stordalen, Sweden	68°21'N, 19°02'E	Subarctic mire	<i>Eriophorum angustifolium</i>	2004–2006	Automatic chamber and eddy covariance	Petrescu et al. (2008)
5	Degero Stormyr, Sweden	64°11'N, 19°33'E	Mixed acid mire	<i>Eriophorum vaginatum</i> , <i>Vaccinium oxycoccus</i> , <i>Andromeda polifolia</i> , and <i>Scheuchzeria palustris</i>	1995–1997	Static chamber	Granberg et al. (2001)
6	Lena Samoylov Delta, Russia	72°22'N, 126°30'E	Tundra	Sedges, mosses, and shrubs	2003–2004	Eddy covariance	Wille et al. (2008)
7	Bakchar, Russia	56°51'N, 82°50'E	Open unforested mesotrophic fen	<i>Eriophorum vaginatum</i> , <i>Carex rostrata</i> , <i>Carex limosa</i> , and horse-tail <i>Equisetum fluviatile</i>	1999	Static chamber	Friborg et al. (2003)
8	Manitoba, Canada	58°40'N, 93°50'W	Eutrophic fen	Sedges, grasses, and moss	2008–2011	Eddy covariance	Hanis et al. (2013)
9	Athabasca, Canada	54°57'N, 112°28'W	Moderately rich treed fen	Stunted trees, shrub, and mosses	2007	Eddy covariance	Long et al. (2010)
10	Quebec, Canada	53°40'N, 78°10'W	Boreal bog	Lichen, mosses, and shrubs	2012	Eddy covariance	Nadeau et al. (2013)
11	Robinsons, Canada	48°16'N, 58°40'W	Boreal bog	Mosses, lichens, and sedge	2014–2016	Eddy covariance	Wang et al. (2018)

**Table 3**  
Description of Sites in Temperate Regions

No.	Site name	Location	Wetland type	Dominant vegetation	Time	Method	References
1	Daxinganling, China	51°7.92'N, 125°8.22'E	Intermittently inundated marsh	<i>Carex meyerian</i> (#1) <sup>a</sup>	2011–2012	Static chamber	Liu et al. (2015)
		52°56'N, 122°51'E	Permafrost peatland	Shrubs, sedges, and grass (#2)	2012–2013	Eddy covariance	Sun et al. (2018)
2	Sanjiang Plain, China	47°53'N, 133°30'E	Permanently inundated marsh	<i>Glyceria spiculosa</i> (#1) and <i>Deyeuxia angustifolia</i> (#2)	2011	Static chamber	Sun et al. (2013)
				<i>Carex lasiocarpa</i> , <i>Glyceria spiculosa</i> , and <i>Deyeuxia angustifolia</i> (#3)	2011	Eddy covariance	
		47°35'N, 133°30'E	Permanently inundated marsh	<i>Carex lasiocarpa</i> and <i>Carex pseudocuraica</i> (#4)	2012–2013	Eddy covariance	Sun et al. (2018)
3	Luanhaizi, China	37°35'N, 101°20'E	Alpine wetland	<i>Carex pamirensis</i> and <i>Carex alrofusca</i>	2012–2013	Eddy covariance	Song et al. (2015)
4	Ruoergai, China	32°47'N, 102°32'E	Peatland	<i>Carex multiensis</i> (#1) and <i>Carex meyeriana</i> (#2)	2001	Static chamber	Ding et al. (2004)
5	Minnesota, United States	47°32'N, 93°28'W	Poor fen	Sphagnum, Carex, and <i>Scheuchzeria palustris</i> (#1)	1988–1990	Static chamber	Dise (1993)
		47°30'N, 93°29'W	Peatland	<i>Sphagnum papillosum</i> , <i>Carex</i> spp., <i>Eriophorum chamissonis</i> , and <i>Saracenia purpurea</i> (#2)	2009–2011	Eddy covariance	Olson et al. (2013)
6	Sallies Fen, United States	43°12.5'N, 71°3.5'W	Poor fen	Mosses, shrubs, and sedges	2008–2011	Static chamber	Noyce et al. (2014)
7	Michigan, United States	42°27'N, 84°01'W	Ombrotrophic peatland	Sphagnum and <i>Scheuchzeria palustris</i>	1991–1993	Static chamber	Shannon and White (1994)
8	Mer Bleue, Canada	45°41'N, 75°52'W	Temperate ombrotrophic bog	Chamaedaphne (#1), <i>Eriophorum</i> (#2), and <i>Maianthemum</i> (#3)	2009–2010	Triplicate auto chamber	Lai et al. (2014)
				Sphagnum spp., Ericaceous shrubs, and Judd. Sedges (#4)	2011–2012	Eddy covariance	Brown et al. (2014)
9	Federseemoor, Germany	48°06'N, 9°38'E	Minerotrophic peatland	Phragmites	2013	Eddy covariance	van den Berg et al. (2016)
10	Schechenfilz, Germany	47°48'N, 11°19'E	Temperate bog-pine forest	Bog-pines and mosses	2012–2013	Eddy covariance	Hommelberg et al. (2014)
11	Kopytkowo, Poland	53°35'N, 22°53'E	Temperate mire	Reeds, sedges, and rushes	2013–2014	Eddy covariance	Fortuniak et al. (2017)
12	Kopuatai, New Zealand	37°56'S, 175°22'E	Ombrotrophic, raised bog	Wire rush and <i>Empodisma robustum</i>	2012–2014	Eddy covariance	Goodrich et al. (2015)

<sup>a</sup>The observed site with corresponding dominant vegetation.

**Table 4**  
Description of Sites in Tropical Regions

No.	Site name	Location	Wetland type	Dominant vegetation	Time	Method	References
1	Sarawak, Malaysia	1°25'N, 111°07'E	Tropical peat swamp forest	<i>Gonystylus bancanus</i> , <i>Dactylocladus stenostachys</i> , and <i>Copaifera palustris</i> (#1)	2012–2015	Closed chamber	Sangok et al. (2017)
		1°27'N, 111°09'E	Tropical peat swamp forest	<i>Shorea albida</i> , <i>Gonystylus bancanus</i> , and <i>Stemonurus</i> spp. (#2)	2013	Eddy covariance	Tang et al. (2018)
				<i>Shorea albida</i> , <i>Lithocarpus</i> sp., <i>Litsea</i> sp., and <i>Dillenia</i> sp. (#3)	2014–2015	Eddy covariance	Wong et al. (2018)

$$Y = M(X) = \sum_{\alpha \in \mathbb{N}^k} \lambda_{\alpha} \Psi_{\alpha}(X). \quad (35)$$

Here, the  $\Psi_{\alpha}(X)$  are multivariate polynomials orthonormal with regard to the distribution of  $X$ ,  $\alpha \in \mathbb{N}^k$  is an indices vector that identifies the components of the multivariate polynomials  $\Psi_{\alpha}$ , and the  $\lambda_{\alpha} \in \mathbb{R}$  represent the corresponding coefficients.

The sum in equation 35 needs to be truncated to a finite sum, when applied in realistic situation, by the truncated PCE.

$$M(X) \approx M^{PC}(X) = \sum_{\alpha \in A} \lambda_{\alpha} \Psi_{\alpha}(X) \quad (36)$$

Here,  $A \in \mathbb{N}^k$  is the set of selected multi-indices of multivariate polynomials.

Therefore, when the model  $M(X)$  is approximated by the PCE surrogate, the Sobol indices can be computed by its coefficients  $\lambda_{\alpha}$  as follows:

$$E(M(X)) \approx \lambda_0, \quad (37)$$

$$V(M(X)) \approx \sum_{\substack{\alpha \in A \\ \alpha \neq 0}} \lambda_{\alpha}^2, \quad (38)$$

$$S_i = \frac{1}{V(M(X))} \sum_{\alpha \in A_{S_i}} \lambda_{\alpha}^2 \text{ with } A_{S_i} = \{\alpha: \alpha_i > 0, \alpha_k = 0 \text{ for } k \neq i\}, \quad (39)$$

$$S_{Ti} = \frac{1}{V(M(X))} \sum_{\alpha \in A_{S_{Ti}}} \lambda_{\alpha}^2 \text{ with } A_{S_{Ti}} = \{\alpha: \alpha_i > 0\}, \quad (40)$$

$$S_{ij} = \frac{1}{V(M(X))} \sum_{\alpha \in A_{S_{ij}}} \lambda_{\alpha}^2 \text{ with } A_{S_{ij}} = \{\alpha: \alpha_i > 0, \alpha_j > 0, \alpha_k = 0 \text{ for } k \neq i, j\}. \quad (41)$$

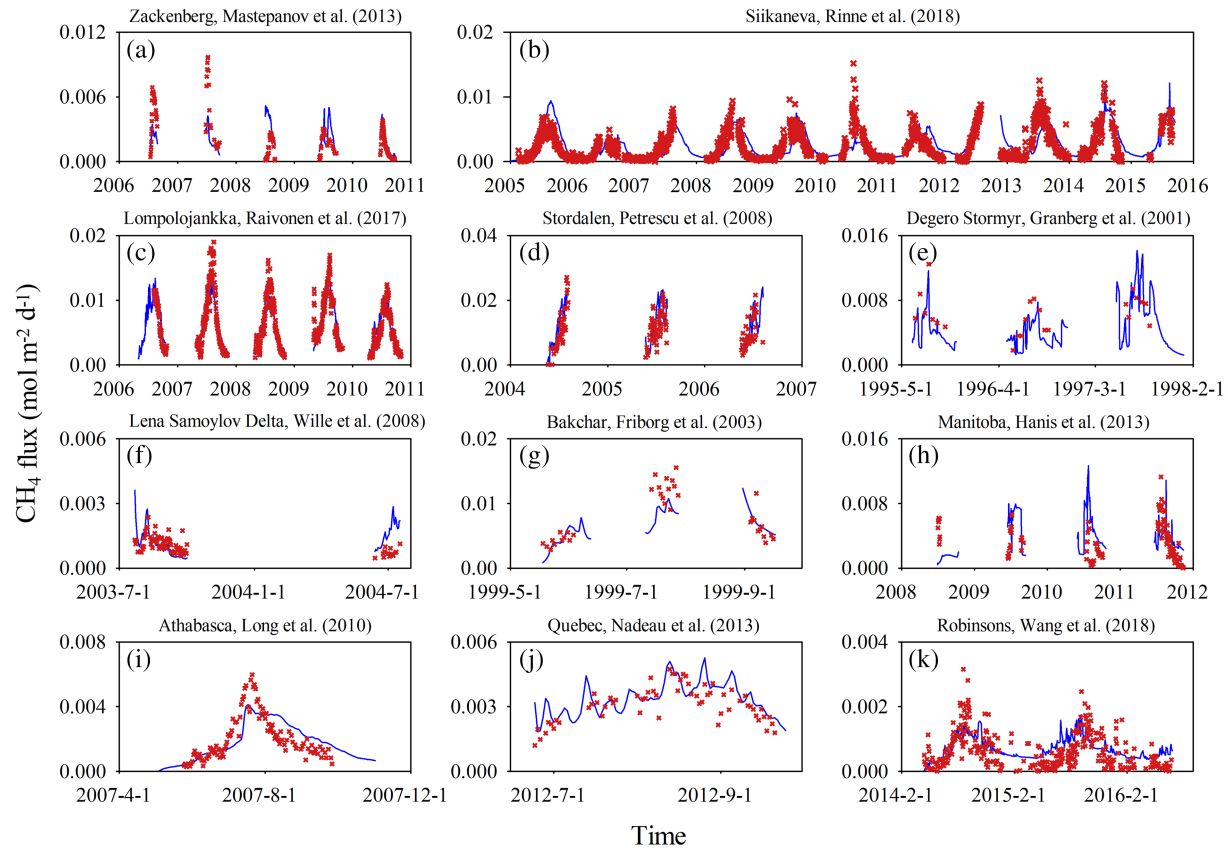
Here,  $E(M(X))$  and  $V(M(X))$  indicate the expectation and variance of  $M(X)$ .  $S_i$ ,  $S_{Ti}$ , and  $S_{ij}$  are the first-order Sobol indices, the total Sobol indices, and the second-order Sobol indices (also called the joint Sobol indices), respectively.  $A_{S_i}$ ,  $A_{S_{Ti}}$ , and  $A_{S_{ij}}$  are the sets of indices vectors that only include the interest term corresponding to the sensitivity index.

The first-order Sobol indices measure the fraction of the variance contributed by the  $i$ th parameter only; the total Sobol indices measure the total variance contribution due to the  $i$ th parameter and its interactions with other parameters; and the second-order Sobol indices measure the fractional variance contribution corresponding to the joint  $i$ th and  $j$ th parameter (Ricciuto et al., 2018).

### 3.4. Model Evaluation

Three metrics were used to evaluate model performance, including:

1. The coefficient of determination ( $R^2$ ), which represents the variation in the observations interpreted by the model.
2. Root mean square error (RMSE), calculated as



**Figure 2.** Daily variation in simulated and observed CH<sub>4</sub> emissions at boreal sites. The blue solid lines represent simulated values, and the red cross dots represent observed values.

$$\text{RMSE} = \sqrt{\frac{\sum_{i=1}^n (M_i - O_i)^2}{n}}. \quad (42)$$

Here,  $M_i$  is the modeled value and  $O_i$  the observed value;  $n$  is the number of days for which we have the observed values.

3. Relative predictive error (RPE), computed as

$$\text{RPE} = \frac{\overline{M} - \overline{O}}{\overline{O}} \times 100\%. \quad (43)$$

Here,  $\overline{M}$  and  $\overline{O}$  are the means of modeled and observed values, respectively.

## 4. Results

### 4.1. Model Evaluation

Twenty-four sites were selected to evaluate the CH<sub>4</sub> model (Tables 2–4). These evaluation sites spanned from boreal to tropical regions and covered several dominant wetland types including fen, bog, mire, marsh, peatland swamp, and tundra. We examined model performance by assessing its ability to accurately reproduce the magnitude and temporal variability of CH<sub>4</sub> emissions among these sites.

#### 4.1.1. Boreal Sites

Observations from 11 sites in boreal regions, mainly located in northern Europe and Canada, were compiled for model evaluation, comprising six fen sites, two bog sites, two mire sites, and one tundra site. Overall, the model reliably predicted the observations (Figure 2), with the RPE varying from −19.47% to 33.28% (Table 5).



**Table 5**  
Statistic Results for Site-Level Evaluation

Site	$R^2$	RMSE <sup>a</sup>	RPE (%)	N
<b>Boreal sites</b>				
Zackenberg	0.210	0.0019	−1.39	192
Siikaneva	0.517	0.0017	14.54	2285
Lompolojännkä	0.815	0.0018	2.80	837
Stordalen	0.317	0.0057	17.38	126
Degero Stormyr	0.361	0.0028	−19.47	23
Lena Samoylov Delta	0.053	0.0007	15.95	79
Bakchar	0.586	0.0028	−16.55	35
Manitoba	0.012	0.0030	30.08	127
Athabasca	0.563	0.0010	4.38	103
Quebec	0.455	0.0007	10.76	61
Robinsons	0.313	0.0006	33.28	329
<b>Temperate sites</b>				
Daxinganling1	0.297	0.0005	12.31	38
Daxinganling2	0.272	0.0002	15.38	242
Sanjiang1	0.258	0.0030	13.46	54
Sanjiang2	0.000	0.0025	−2.79	57
Sanjiang3	0.379	0.0032	−6.27	119
Sanjiang4	0.413	0.0027	7.04	142
Luanhaizi	0.170	0.0025	−26.70	133
Ruoergai1	0.118	0.0029	20.29	35
Ruoergai2	0.059	0.0039	−19.26	35
Minnesota1	0.689	0.0057	−9.13	54
Minnesota2	0.506	0.0033	4.22	644
Sallies Fen	0.240	0.0052	−28.70	80
Michigan	0.142	0.0147	−27.30	62
Mer Bleue1	0.083	0.0018	−0.19	194
Mer Bleue2	0.498	0.0034	−13.63	208
Mer Bleue3	0.231	0.0028	−28.99	185
Mer Bleue4	0.000	0.0012	−22.68	276
Federseemoor	0.252	0.0043	−3.55	253
Schechenfilz	0.231	0.0012	32.64	145
Kopytkowo	0.145	0.0060	−48.62	239
Kopuatai	0.001	0.0041	−4.97	124
<b>Tropical sites</b>				
Sarawak1	0.016	0.0003	47.04	33
Sarawak2	0.003	0.0010	25.08	61
Sarawak3	0.426	0.0005	6.25	18

<sup>a</sup>The unit of RMSE is  $\text{mol m}^{-2} \text{d}^{-1}$ .

The magnitude and seasonality of simulated  $\text{CH}_4$  emissions were consistent with the observations, especially at the Siikaneva and Lompolojännkä sites (Figures 2b and 2c). Long-term continuous  $\text{CH}_4$  observations had been collected at these two sites, and the values of the coefficient of determination ( $R^2$ ) were about 0.52 and 0.82, respectively (Table 5).

Although the model explained variability in  $\text{CH}_4$  emissions well at most sites, large differences between simulated and observed  $\text{CH}_4$  emissions were still observed to exist. The model underestimated  $\text{CH}_4$  observations at the Zackenberg site; the discrepancies between modeled and observed values mainly occurred in the growing seasons of 2006 and 2007 (Figure 2a). Meanwhile, the model had slight high predicted values in the growing season of 2004 at Lena Samoylov Delta (Figure 2f). At these two sites,  $R^2$  was very low with values of 0.21 and 0.053, respectively (Table 5). In addition, the model predicted slightly lower values for peak emissions at the Bakchar, Athabasca, and Robinsons sites (Figures 2g, 2i, and 2k). The lowest  $R^2$  was produced at the Manitoba site, which simultaneously had a high RPE value ( $R^2 = 0.012$ ,  $\text{RPE} = 30.08\%$ ) (Table 5), indicating that the model failed to capture the variation and magnitude of  $\text{CH}_4$  observations at this site.

#### 4.1.2. Temperate Sites

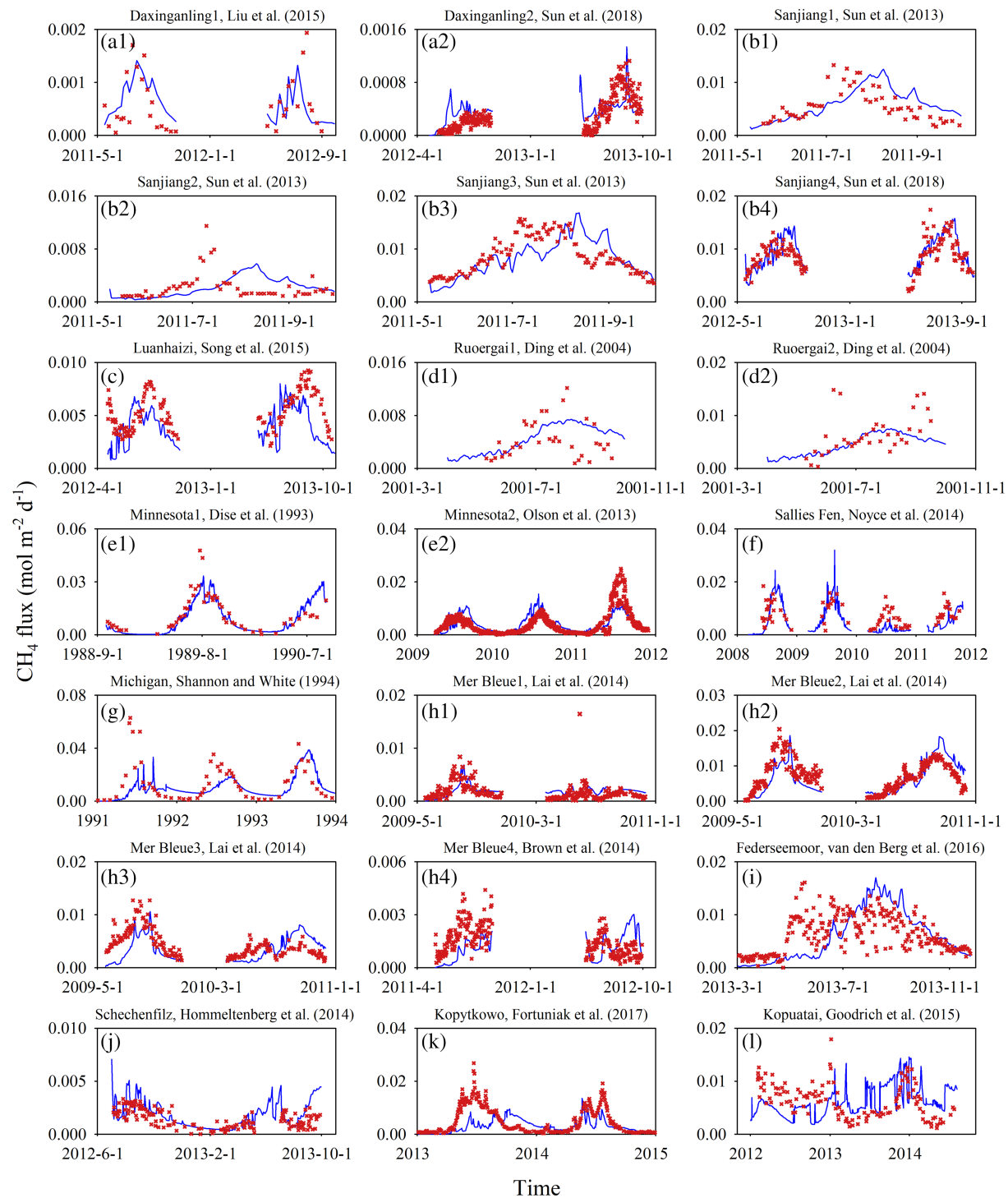
$\text{CH}_4$  flux observations from 12 sites in temperate regions were compiled for model evaluation. On the whole, the model predictions matched the observations well (Figure 3). At the Minnesota and Mer Bleue sites (Figures 3e1, 3e2, and 3h2), the model explained about 69%, 51%, and 50% of the variation observed in  $\text{CH}_4$  emissions, respectively (Table 5). The RPE ranged from −28.99% to 32.64%, except at the Kopytkowo site, demonstrating that a lower deviation between simulated and observed  $\text{CH}_4$  emissions existed at most sites (Table 5).

However, large differences also existed between modeled and observed  $\text{CH}_4$  emissions at some sites. At the Sanjiang sites, the simulated peak emissions clearly lagged the observed peak values (Figures 3b1–3b3). Although they had lower RPEs, with values of 13.46%, −2.79%, and −6.27%, respectively (Table 5), the model predicted little of the variability in observed  $\text{CH}_4$  emissions, with  $R^2$  being 0.258, 0.000, and 0.379, respectively (Table 5). At the Ruoergai and Federseemoor sites, observed  $\text{CH}_4$

fluxes had no general seasonal patterns (Figures 3d1, 3d2, and 3i). The  $R^2$  values were 0.118, 0.059, and 0.252, and the RMSE values were 0.0029, 0.0039, and 0.0043  $\text{mol m}^{-2} \text{d}^{-1}$ , respectively (Table 5). Thus, the simulated results only matched the magnitude of observed  $\text{CH}_4$  emissions at these two sites. The model underestimated observed  $\text{CH}_4$  emissions at the Daxinganling, Minnesota, Michigan and Mer Bleue sites (Figures 3a2, 3e2, 3g, and 3h4), which mainly occurred in the 2013 growing season at Daxinganling, 2011 at Minnesota, 1991 at Michigan, and 2011 at Mer Bleue. Meanwhile, the model slightly overestimated  $\text{CH}_4$  emissions at Daxinganling and Mer Bleue in 2012 (Figures 3a2 and 3h4). In addition, the model showed poor agreement between simulated and observed  $\text{CH}_4$  emissions at the Luanhaizi, Kopytkowo, and Kopuatai sites (Figures 3c, 3k, and 3l). The calculated  $R^2$  and RPE values at these sites were 0.17 and −26.70%, 0.145 and −48.62%, and 0.001 and −4.97%, respectively (Table 5). At the Kopytkowo site, the model significantly underestimated observations during the growing season in 2013 (Figure 3k).

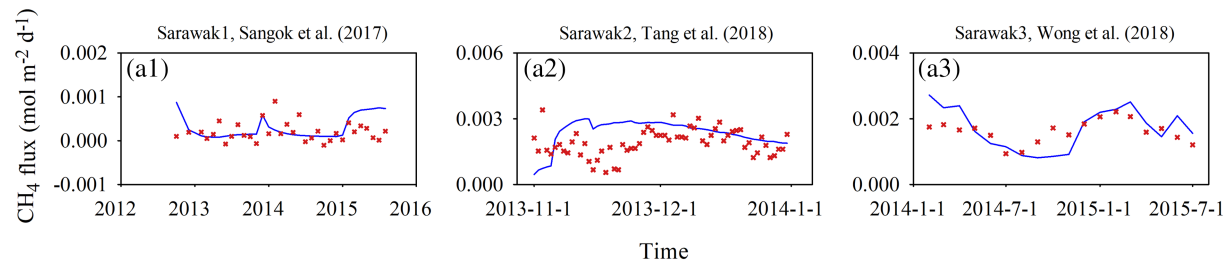
#### 4.1.3. Tropical Sites

Observations from a tropical peat swamp forest located in Sarawak, Malaysia were also collected for comparison. Overall, the model showed good agreement between simulated  $\text{CH}_4$  emissions and observations at this site (Figure 4). At the Sarawak1 site, the simulated  $\text{CH}_4$  emissions did not reproduce the variation in observations, which had no clear seasonal patterns (Figure 4a1). In addition, the model slightly overestimated the



**Figure 3.** Daily variation in simulated and observed  $\text{CH}_4$  emissions at temperate sites. The blue solid lines represent simulated values, and the red cross dots represent observed values.

observed  $\text{CH}_4$  emissions in November at the Sarawak2 site (Figure 4a2). The  $R^2$  and RPE values for these two sites were 0.016 and 47.04% and 0.003 and 25.08%, respectively (Table 5). The model captured the magnitude and variation of  $\text{CH}_4$  emissions during the measured period at the Sarawak3 site (Figure 4a3). The  $R^2$  and RPE values for this site were 0.426 and 6.25%, respectively (Table 5). The RMSE values for this peat



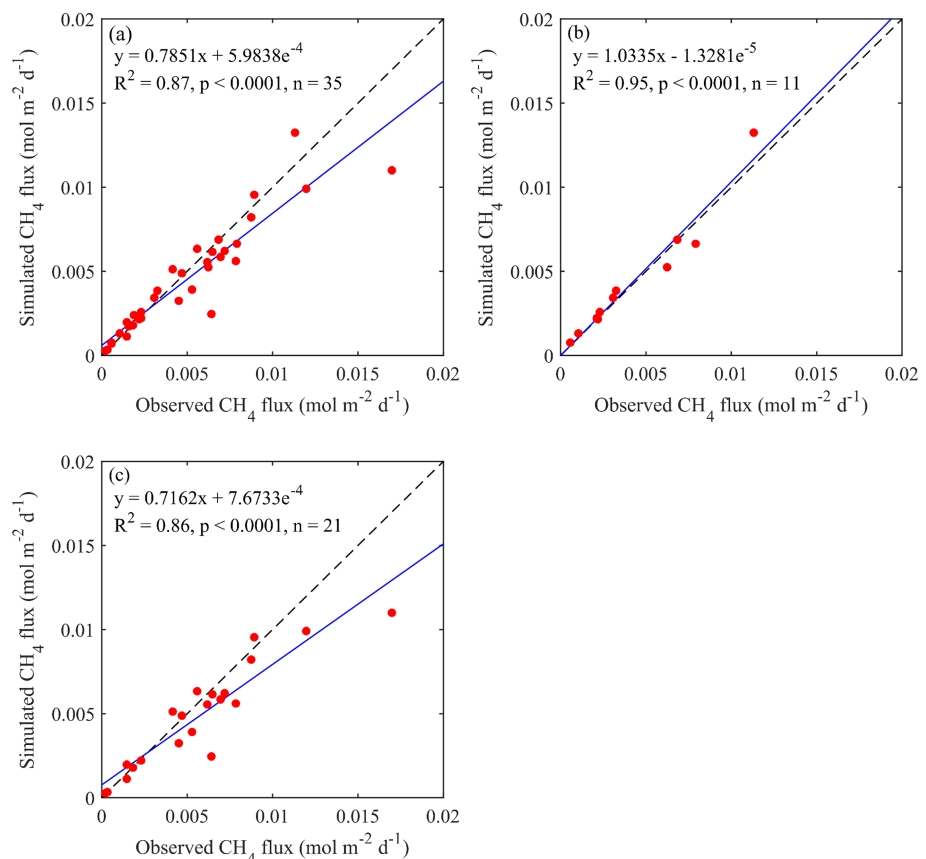
**Figure 4.** Daily variation in simulated and observed  $\text{CH}_4$  emissions at a tropical site. The blue solid lines represent simulated values, and the red cross dots represent observed values.

swamp forest site ranged from 0.0003 to 0.001  $\text{mol m}^{-2} \text{d}^{-1}$  (Table 5), indicating a small disparity between simulated and observed  $\text{CH}_4$  emissions.

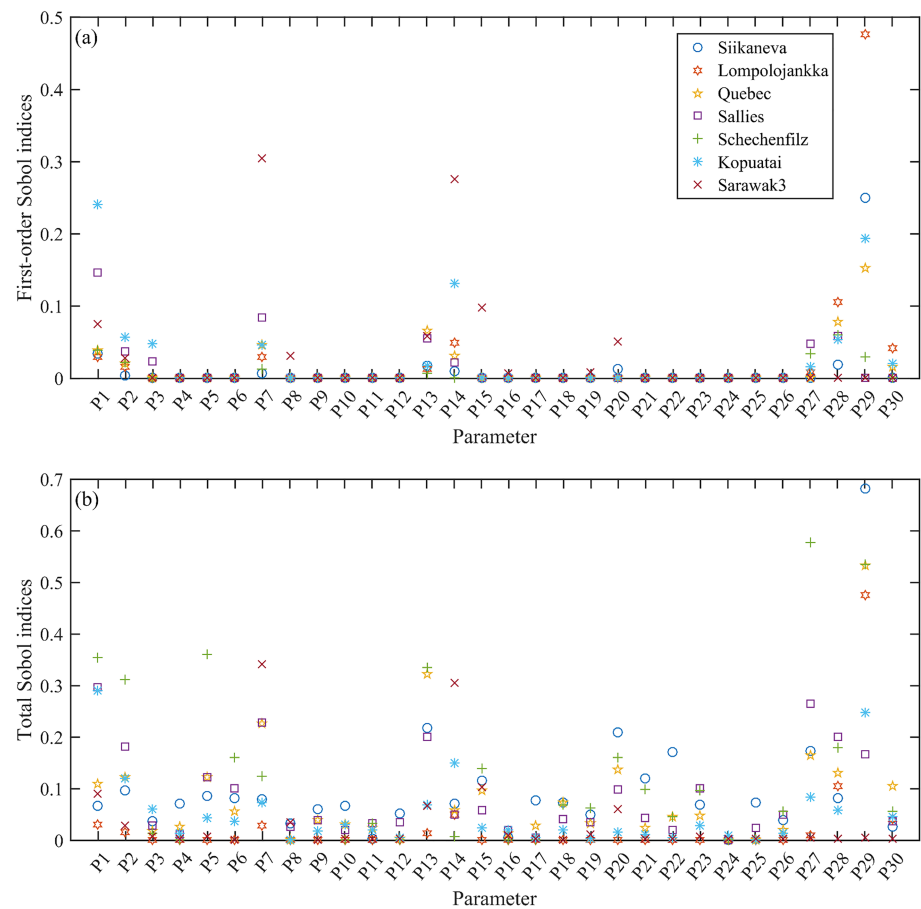
The average simulated and observed  $\text{CH}_4$  emissions during measured periods for all evaluated, boreal and temperate sites are shown in Figure 5. On the whole, the average values of simulations and observations at all evaluated sites lie close to the 1:1 line over a range of 0 to 0.017  $\text{mol m}^{-2} \text{d}^{-1}$ , and  $R^2$  is 0.87 (Figure 5a).

#### 4.2. Sensitivity Analysis

The PCE-based parameter sensitivity analysis of modeled  $\text{CH}_4$  emissions with 500 model evaluations was performed at seven measured sites, covering the tropical, temperate, and boreal biomes (Figures 6 and 7). For the Siikaneva, Lompolojännkä, and Quebec sites,  $\text{CH}_4\text{prodQ}_{10}$  (P29) is the most sensitive parameter for  $\text{CH}_4$  emissions (Figure 6a), which is also reflected in the evaluation of total Sobol indices (Figure 6b).



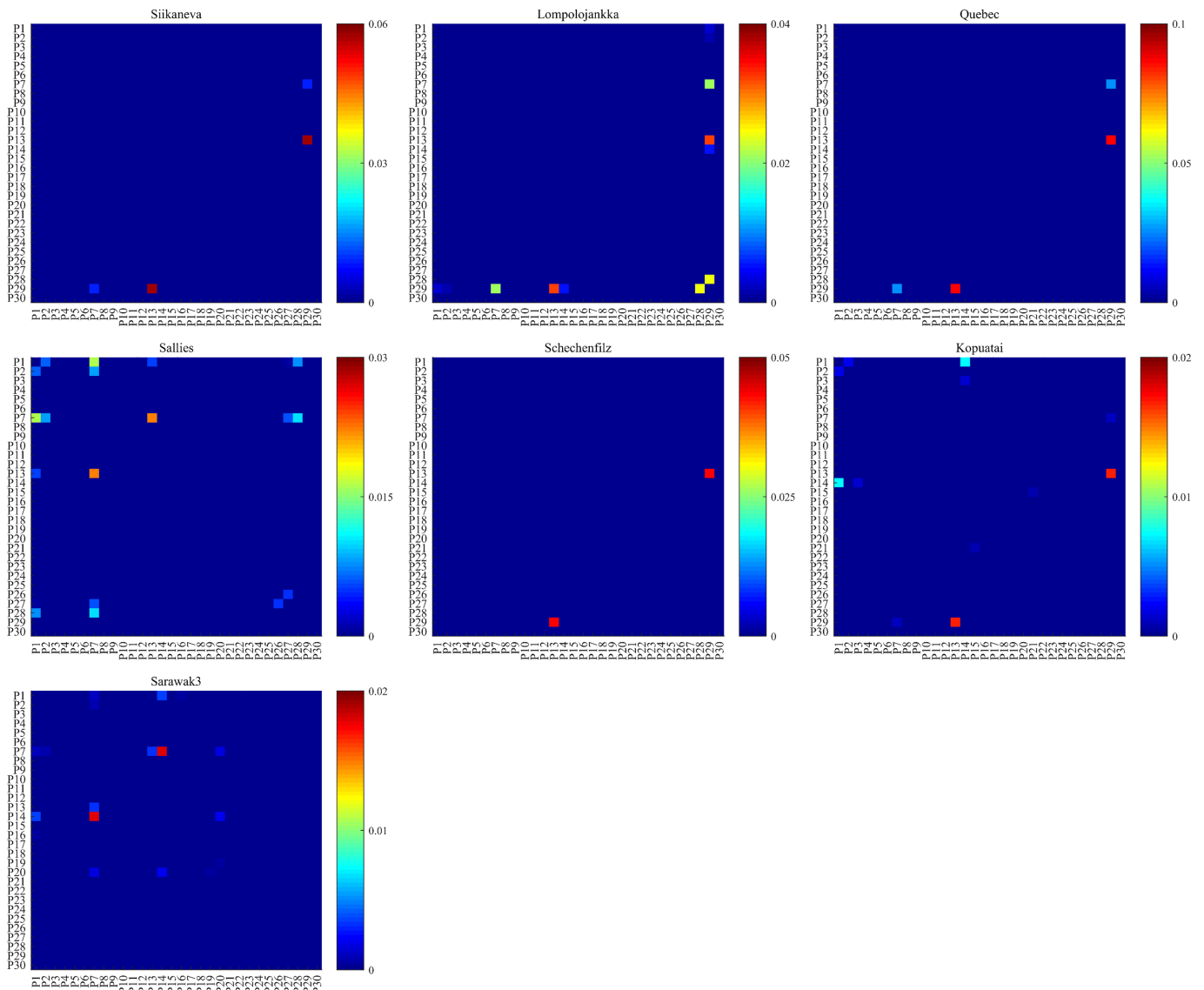
**Figure 5.** Comparison of the average observed and simulated  $\text{CH}_4$  emissions (red dots). (a) All evaluated sites, (b) boreal sites, and (c) temperate sites.



**Figure 6.** Parameter sensitivity indices for modeled CH<sub>4</sub> emissions. (a) First-order Sobol indices and (b) total Sobol indices. All parameters are listed in Table 1.

For the Sallies and Kopuatai sites,  $K_{\text{cpool}}$  (P1) has the largest impact on CH<sub>4</sub> emissions (Figures 6a and 6b). In addition, The CH<sub>4</sub> emissions are also sensitive to  $V_{\text{Acecons,max}}$  (P7),  $K_{\text{AceprodCH}_4}$  (P13), and  $\text{DOCprodQ}_{10}$  (P27) at the Sallies site and sensitive to  $K_{\text{CH}_4\text{prod}}$  (P14) and  $\text{CH}_4\text{prodQ}_{10}$  at the Kopuatai site (Figures 6a and 6b). For the Schechenfilz site, the first-order Sobol indices are low for all parameters (Figure 6a). However, the  $\text{DOCprodQ}_{10}$  has a great total effect index of about 58% on the total variance of CH<sub>4</sub> emissions (Figure 6b), which indicates the apparent impact of this parameter on the CH<sub>4</sub> emissions. For the Sarawak3 site, except for  $V_{\text{Acecons,max}}$  and  $K_{\text{CH}_4\text{prod}}$ , the first-order and total Sobol indices have low values for other parameters (Figures 6a and 6b). CH<sub>4</sub> emissions at this site are obviously sensitive to these two parameters, with  $V_{\text{Acecons,max}}$  becoming the most sensitive parameter (Figures 6a and 6b). Moreover, compared to other sites, the Sobol indices of parameters related to temperature sensitivity (P27–P30) are very low at the Sarawak3 site, which may be related to the high and relatively constant soil temperature in the tropics.

The second-order Sobol indices results indicate that  $K_{\text{AceprodCH}_4}$  and  $\text{CH}_4\text{prodQ}_{10}$  have the largest interaction effects on CH<sub>4</sub> emissions at the Siikaneva, Lompolojankkä, Quebec, Schechenfilz, and Kopuatai sites (Figure 7). The parameters  $V_{\text{Acecons,max}}$  and  $K_{\text{AceprodCH}_4}$  have the largest interaction effects on CH<sub>4</sub> emissions at the Sallies site, and  $V_{\text{Acecons,max}}$  and  $K_{\text{CH}_4\text{prod}}$  have the largest interaction effects on CH<sub>4</sub> emissions at the Sarawak3 site (Figure 7). In addition,  $V_{\text{Acecons,max}}$  and  $\text{CH}_4\text{prodQ}_{10}$  have the important interaction effects on CH<sub>4</sub> emissions at the Siikaneva, Lompolojankkä, and Quebec sites (Figure 7). For the Lompolojankkä site, the parameter subset of  $\text{AceprodQ}_{10}$  and  $\text{CH}_4\text{prodQ}_{10}$  also has the important interaction effects on CH<sub>4</sub> emissions (Figure 7). The parameter set of  $K_{\text{cpool}}$  and  $V_{\text{Acecons,max}}$  and the parameter set of  $V_{\text{Acecons,max}}$  and  $\text{AceprodQ}_{10}$  also have the important interaction effects on CH<sub>4</sub> emissions at the Sallies site (Figure 7).



**Figure 7.** The second-order Sobol indices for modeled  $\text{CH}_4$  emissions. All parameters are listed in Table 1.

Although the results of first-order, total, and second-order Sobol indices among the parameters are somewhat different between these sites, the main sensitivity parameters are  $\text{DOCprodQ}_{10}$ ,  $\text{AceprodQ}_{10}$ ,  $\text{CH}_4\text{prodQ}_{10}$ ,  $K_{\text{cpool}}$ ,  $V_{\text{Acecons,max}}$ ,  $K_{\text{AceprodCH}_4}$ , and  $K_{\text{CH}_4\text{prod}}$ , which indicates that these parameters that control DOC and acetate production and acetoclastic methanogenesis have the significant impact on modeled  $\text{CH}_4$  emissions.

### 4.3. Environmental Controls on $\text{CH}_4$ Emissions

At the above test sites, inconsistent relationships between simulated  $\text{CH}_4$  emissions, soil temperature, and measured water table position were observed (Figure S1). For example, the seasonality of simulated  $\text{CH}_4$  emissions at the Siikaneva and Lompolojankka sites depended on soil temperature dynamics and had no significant correlation with water table position. However, at the Sarawak3 site, the temporal variation of simulated  $\text{CH}_4$  emissions coincided well with water table position but showed no dependence on soil temperature. In addition, combined effects of soil temperature and water table position were observed on simulated  $\text{CH}_4$  emissions at other sites. The soil temperature and water table position exerted a positive effect on simulated  $\text{CH}_4$  emissions during the growing season at the Quebec site, where peak emissions





**Figure 8.** Time series of simulated  $\text{CH}_4$  production, oxidation, and transportation.

occurred when soil temperature and water table position reached their maximum. However, although soil temperature had a positive effect at the other sites, an increase of water table position had little influence on simulated  $\text{CH}_4$  emissions, and a decrease substantially reduced  $\text{CH}_4$  emissions, especially during the growing season.

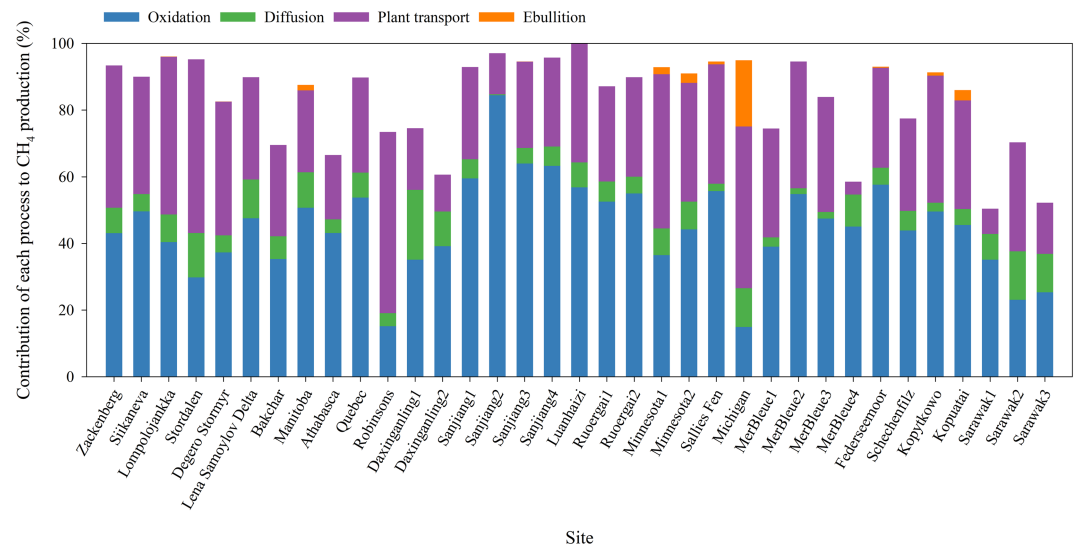
## 5. Discussion

### 5.1. Modeled Process Components

The new  $\text{CH}_4$  model explicitly represented  $\text{CH}_4$ -related microbial mechanisms, including anaerobic fermentation, homoacetogenesis, hydrogenotrophic methanogenesis, acetoclastic methanogenesis, and methanotrophy, and the interaction of different gases, including  $\text{CH}_4$ ,  $\text{O}_2$ ,  $\text{CO}_2$ , and  $\text{H}_2$ , in wetland  $\text{CH}_4$  emission modeling. The new model has also been integrated into a terrestrial ecosystem model (IBIS) and evaluated at 24 different wetland sites globally. The simulated results captured the magnitude and variation of observed  $\text{CH}_4$  emissions at most sites.

Net  $\text{CH}_4$  emission is determined by the  $\text{CH}_4$  production by methanogens,  $\text{CH}_4$  oxidation by methanotrophs, and molecular diffusion, plant-mediated transport, and ebullition to the atmosphere (Chanton, 2005). To assess these processes represented by the model, we calculated the temporal variation in process components of wetland  $\text{CH}_4$  emissions at the test sites (Figure 8) and the proportion that each process relative to  $\text{CH}_4$  production at all evaluated sites (Figure 9).

Methanotrophy is an important process and can be up to 100% of  $\text{CH}_4$  production (Fritz et al., 2011; Whalen, 2005). An almost completely  $\text{CH}_4$  oxidation in a densely rooted bog has been reported recently from an analysis of isotopic  $\delta^{13}\text{C}$ - $\text{CH}_4$  values (e et al., 2019). The  $\text{CH}_4$  oxidation of global wetland has also been estimated to be 40–70% of total  $\text{CH}_4$  production (Meronigal et al., 2004). In our results,  $\text{CH}_4$  production and oxidation showed apparent temporal variation at all test sites, with peak values of both processes occurring in the midgrowing seasons (Figure 8). This may reflect enhanced microbial activities caused by the high



**Figure 9.** Simulated contributions (%) of  $\text{CH}_4$  oxidation, diffusion, plant transport, and ebullition processes to  $\text{CH}_4$  production (which is set to 100%); the rest is left in the soil.

temperature in this period (Dunfield et al., 1993). In addition, at all evaluated sites,  $\text{CH}_4$  oxidation was observed to be about half or more of  $\text{CH}_4$  production (Figure 9).

Plant-mediated transport has been reported as the principal pathway for  $\text{CH}_4$  emissions (Colmer, 2003; Green & Baird, 2012; King et al., 1998; Waddington et al., 1996; Whiting & Chanton, 1992). In our analysis, plant-mediated transport also showed clear temporal variation and played a primary role in total observed period, followed by molecular diffusion, and ebullition had the smallest contribution and being characterized by pulse behavior (Figure 8). Moreover, plant-mediated transport also had the greatest proportion of total  $\text{CH}_4$  emission at most sites (Figure 9).

## 5.2. Environmental Controls on $\text{CH}_4$ Emissions

Numerous studies have identified soil temperature as the fundamental control on wetland  $\text{CH}_4$  emissions (Liu et al., 2015; Rinne et al., 2018; Sun et al., 2018), with an exponential dependence of  $\text{CH}_4$  emissions on soil temperature being recorded (Marushchak et al., 2016; Mikhaylov et al., 2015). Water table position is also generally recognized as a major physical control on  $\text{CH}_4$  emissions from wetlands (Moore et al., 2011; Wong et al., 2018). The relationship between water table position and  $\text{CH}_4$  emissions is usually nonmonotonic (Brown et al., 2014; Christensen et al., 2003). However, several studies have reported no significant dependence of  $\text{CH}_4$  emissions on water table position (Jackowicz-Korczyński et al., 2010; Liu et al., 2015; Rinne et al., 2018). In addition, combined effects of soil temperature and water table position are observed on  $\text{CH}_4$  emissions from wetlands (Goodrich et al., 2015; Lai et al., 2014; Noyce et al., 2014).

This study analyzed the relationships between modeled  $\text{CH}_4$  emissions and soil temperature and observed water table position at seven test sites. The different control patterns of these two variables on observed  $\text{CH}_4$  emissions are reflected in our simulations (see section 3.3). At the Lompolojankka site, the water table position is almost always above the soil surface (Figure S1); this may maintain a relatively stable environment for wetland  $\text{CH}_4$  dynamics. Thus, the water table position had little or no effect on  $\text{CH}_4$  emissions, which was also observed in continuously inundated ecosystems (Strachan et al., 2015; Sturtevant et al., 2016; Sun et al., 2013). The seasonal variation of simulated  $\text{CH}_4$  emissions at the Siikanen site was controlled by soil temperature (Figure S1), which was consistent with the analysis of observations (Rinne et al., 2018). For the opposite control pattern on simulated  $\text{CH}_4$  emissions at the Sarawak3 site (Figure S1), which may be due to the high and narrow range of soil temperature in tropical regions (Hirano et al., 2014; Melling et al., 2005).

### 5.3. Model Limitations

Comparing to site-level observations inevitably leads to differences and occasionally substantial deviations. Our evaluation identified large differences between modeled and observed CH<sub>4</sub> emissions at some sites. An interpretation for the differences between predictions and observations is the inadequacy of model algorithms. Although we used several microbial mechanisms to represent CH<sub>4</sub> production and oxidation, no observations can be acquired at the site level to evaluate these microbial dynamics, which needs to be improved in the future (Allison et al., 2010). Ebullition is an episodic and complex process that depends on the total partial pressure of dissolved gases and atmospheric and hydrostatic pressure (Tokida et al., 2007). Although we adopted the hydrostatic equilibrium-based algorithm to describe this process, the modeled ebullition may also contribute to the differences to reproduce the observed CH<sub>4</sub> emissions, especially for peak values (Wania et al., 2010).

The gridded inputs play an important role in the prediction at regional and global scales, while local environmental conditions may differ significantly, especially for the meteorological conditions (Wania et al., 2010). Deviations in these conditions will be added into the simulation of vegetation production and soil hydrothermal dynamics and so will propagate into the CH<sub>4</sub> biogeochemical modeling (Riley et al., 2011). Moreover, the coarse time resolution of daily meteorological values may not capture the sharp changes accurately in the processes related to CH<sub>4</sub> emissions, like ebullition. Thus, the uncertainties from forcing data may also contribute to the differences between predictions and observations.

The performance of this new model is comparable with other models. Raivonen et al. (2017) developed a CH<sub>4</sub> emission model of peatland soils (Helsinki Model of Methane buiLd-up and emission, HIMMELI) and evaluated at the Siikanen site. Overall, HIMMELI model showed the coefficient of determination ( $R^2$ ) between observations and / of 0.63 at the Siikanen site, which are comparable with our model of 0.52 (Table 5). Wania et al. (2010) evaluated the performance of LPJ-WHyMe (Lund-Potsdam-Jena Wetland Hydrology and Methane) at seven sites, and three of seven sites also are included in our model evaluation (i.e., Michigan, Minnesota, and Ruohogai). The RMSE of LPJ-WHyMe at these three sites range from 0.0011 to 0.014 mol m<sup>-2</sup> d<sup>-1</sup> using global parameters, and the RMSE of our model are quite close with them (0.0029 to 0.0147 mol m<sup>-2</sup> d<sup>-1</sup>) (Table 5).

Continuous observation and modeling work should be conducted to improve our knowledge of wetland CH<sub>4</sub> dynamics. It is necessary to measure multiple gases at different temporal and spatial scales and separate their different transport pathways (Bridgman et al., 2013). For those areas with sparse observations, such as the tropics, more extensive measurements are particularly needed. Soil carbon and hydrothermal dynamics are important for CH<sub>4</sub> biogeochemical modeling (Kaiser et al., 2017). Although both have been integrated into terrestrial ecosystem models, the improvement of these processes, especially the modeling of water table dynamics, should be made at the fine scale (Zhu et al., 2014). In addition, CH<sub>4</sub> production and oxidation driven by microbes, the important predecessor processes of CH<sub>4</sub> emissions, require further breakthroughs in experiments and observations and should be better to serve to the evaluation of the microbial dynamics (Xu et al., 2016).

## 6. Conclusions

A new wetland CH<sub>4</sub> emission model was developed and integrated into a terrestrial ecosystem model (IBIS). The new model fully considered CH<sub>4</sub> production, oxidation, and three transport pathways and the interaction between CH<sub>4</sub> and other gases and used four main microbial mechanisms to represent CH<sub>4</sub> production and oxidation. We evaluated the model at 24 globally representative wetland sites. The simulated and observed results showed good agreement for most sites in terms of emission magnitude and variability, and the mean simulated and observed values were highly correlated with an  $R^2$  of 0.87. Sensitivity analysis indicated that those controlling DOC and acetate production and acetoclastic methanogenesis are the main parameters that affect CH<sub>4</sub> emissions. The new process-based model is an attempt to incorporate the microbial mechanisms into the wetland CH<sub>4</sub> emission modeling. In the future, more complete observations and better integration with terrestrial ecosystem models will help to reduce the uncertainties in prediction.

## Appendix A: Microbial Dynamics Related to CH<sub>4</sub> Production and Oxidation

Microbial dynamics related to CH<sub>4</sub> production and oxidation are represented by the following equations, which are derived from Kettunen (2003) and Grant (1998):

$$\frac{\partial \text{Homoacetogens}}{\partial t} = \text{Homoacetogens}_{\text{growth}} - \text{Homoacetogens}_{\text{death}}, \quad (\text{A1})$$

$$\frac{\partial \text{H2methanogens}}{\partial t} = \text{H2methanogens}_{\text{growth}} - \text{H2methanogens}_{\text{death}}, \quad (\text{A2})$$

$$\frac{\partial \text{Acemethanogens}}{\partial t} = \text{Acemethanogens}_{\text{growth}} - \text{Acemethanogens}_{\text{death}}, \quad (\text{A3})$$

$$\frac{\partial \text{Methanotrophs}}{\partial t} = \text{Methanotrophs}_{\text{growth}} - \text{Methanotrophs}_{\text{death}}, \quad (\text{A4})$$

where

$$\text{Homoacetogens}_{\text{growth}} = \text{Grow}_{\text{Homoacetogens}} \times 4 \times \text{H2prodAce}, \quad (\text{A5})$$

$$\text{H2methanogens}_{\text{growth}} = \text{Grow}_{\text{H2methanogens}} \times 4 \times \text{H2prodCH4}, \quad (\text{A6})$$

$$\text{Acemethanogens}_{\text{growth}} = \text{Grow}_{\text{Acemethanogens}} \times \text{Acecons}, \quad (\text{A7})$$

$$\text{Methanotrophs}_{\text{growth}} = \text{Grow}_{\text{Methanotrophs}} \times \text{R}_{\text{oxid,CH}_4}, \quad (\text{A8})$$

$$\text{Homoacetogens}_{\text{death}} = \text{Dead}_{\text{Homoacetogens}} \times \text{Homoacetogens} \times f_{T1}, \quad (\text{A9})$$

$$\text{H2methanogens}_{\text{death}} = \text{Dead}_{\text{H2methanogens}} \times \text{H2methanogens} \times f_{T2}, \quad (\text{A10})$$

$$\text{Acemethanogens}_{\text{death}} = \text{Dead}_{\text{Acemethanogens}} \times \text{Acemethanogens} \times f_T(\text{CH4prodQ}_{10}), \quad (\text{A11})$$

$$\text{Methanotrophs}_{\text{death}} = \text{Dead}_{\text{Methanotrophs}} \times \text{Methanotrophs} \times f_T(\text{CH4oxidQ}_{10}). \quad (\text{A12})$$

Here,  $\text{Homoacetogens}_{\text{growth}}$  and  $\text{Homoacetogens}_{\text{death}}$  are the growth and the death of homoacetogens, respectively;  $\text{H2methanogens}_{\text{growth}}$  and  $\text{H2methanogens}_{\text{death}}$  are the growth and death of hydrogenotrophic methanogens, respectively;  $\text{Acemethanogens}_{\text{growth}}$  and  $\text{Acemethanogens}_{\text{death}}$  are the growth and the death of acetoclastic methanogens, respectively; and  $\text{Methanotrophs}_{\text{growth}}$  and  $\text{Methanotrophs}_{\text{death}}$  are the growth and the death of methanotrophs, respectively.  $\text{Grow}_{\text{Homoacetogens}}$ ,  $\text{Grow}_{\text{H2methanogens}}$ ,  $\text{Grow}_{\text{Acemethanogens}}$ , and  $\text{Grow}_{\text{Methanotrophs}}$  are the growth efficiency for the corresponding microbial functional group, the values of which are 0.2, 0.2, 0.3, and 0.4, respectively;  $\text{Dead}_{\text{Homoacetogens}}$ ,  $\text{Dead}_{\text{H2methanogens}}$ ,  $\text{Dead}_{\text{Acemethanogens}}$  and  $\text{Dead}_{\text{Methanotrophs}}$  are the death rate ( $\text{d}^{-1}$ ) for the corresponding microbial functional group, whose values are uniformly set to 0.06.

## Appendix B: Coefficients Associated With Transportation of Gases.

The diffusivities of four gases in air and water (Tang et al., 2010) are calculated as

$$D_{\text{CH}_4}^{\text{air}} = 1.9 \times 10^{-5} \times \left( \frac{T}{T_0} \right)^{1.82}, \quad (\text{B1})$$

$$D_{\text{O}_2}^{\text{air}} = 1.8 \times 10^{-5} \times \left( \frac{T}{T_0} \right)^{1.82}, \quad (\text{B2})$$

$$D_{\text{CO}_2}^{\text{air}} = 1.47 \times 10^{-5} \times \left( \frac{T}{T_0} \right)^{1.792}, \quad (\text{B3})$$

$$D_{\text{H}_2}^{\text{air}} = 6.68 \times 10^{-5} \times \left( \frac{T}{T_0} \right)^{1.82}, \quad (\text{B4})$$

$$D_{\text{CH}_4}^{\text{water}} = 1.5 \times 10^{-9} \times \left( \frac{T}{T_\theta} \right), \quad (\text{B5})$$

$$D_{\text{O}_2}^{\text{water}} = 2.4 \times 10^{-9} \times \left( \frac{T}{T_\theta} \right), \quad (\text{B6})$$

$$D_{\text{CO}_2}^{\text{water}} = 1.81 \times 10^{-6} \times \exp\left(\frac{-2,032.6}{T}\right), \quad (\text{B7})$$

$$D_{\text{H}_2}^{\text{water}} = 5.11 \times 10^{-9} \times \left( \frac{T}{T_\theta} \right). \quad (\text{B8})$$

The Henry's law constant for each gas is computed following Sander (2015), which can be expressed as

$$H_{\text{CH}_4} = 1.3 \times 10^{-3} \times \exp\left[1,700 \times \left(\frac{1}{T} - \frac{1}{T_\theta}\right)\right], \quad (\text{B9})$$

$$H_{\text{O}_2} = 1.3 \times 10^{-3} \times \exp\left[1,500 \times \left(\frac{1}{T} - \frac{1}{T_\theta}\right)\right], \quad (\text{B10})$$

$$H_{\text{CO}_2} = 3.4 \times 10^{-2} \times \exp\left[2,400 \times \left(\frac{1}{T} - \frac{1}{T_\theta}\right)\right], \quad (\text{B11})$$

$$H_{\text{H}_2} = 7.8 \times 10^{-4} \times \exp\left[530 \times \left(\frac{1}{T} - \frac{1}{T_\theta}\right)\right]. \quad (\text{B12})$$

Here,  $T$  is the soil temperature (K), and  $T_\theta$  and  $T_\theta$  are reference temperatures (K) with values of 273.15 and 298, respectively. In here,  $H_X$  is expressed in  $\text{M atm}^{-1}$  and can be converted to  $\text{mol m}^{-3} \text{ Pa}$  by multiplying with the conversion factor  $\theta$ , which value is  $9.8623 \times 10^{-3}$  (Sander, 2015).

The Schmidt number for each gas is calculated following Wania et al. (2010), which can be expressed as

$$\text{Sc}_{\text{CH}_4} = 1,898 - 110.1 \times T_s + 2.834 \times T_s^2 - 0.02791 \times T_s^3, \quad (\text{B13})$$

$$\text{Sc}_{\text{O}_2} = 1,800.6 - 120.1 \times T_s + 3.7818 \times T_s^2 - 0.047608 \times T_s^3, \quad (\text{B14})$$

$$\text{Sc}_{\text{CO}_2} = 1,911 - 113.7 \times T_s + 2.967 \times T_s^2 - 0.02943 \times T_s^3, \quad (\text{B15})$$

$$\text{Sc}_{\text{H}_2} = 629.95 - 34.691 \times T_s + 0.8681 \times T_s^2 - 0.0084 \times T_s^3. \quad (\text{B16})$$

Here,  $T_s$  is the soil temperature in degree Celsius.

### Appendix C: Environmental Controls.

The soil temperature factors are calculated as

$$f_T(Q_{10}) = \begin{cases} 0 & T < 0 \\ Q_{10}^{\frac{T-30}{10}} & 0 \leq T \leq 30 \\ 1 & T > 30 \end{cases}, \quad (\text{C1})$$

$$f_{T1} = \frac{(T - T_{\min1}) \times (T - T_{\max1})}{(T - T_{\min1}) \times (T - T_{\max1}) - (T - T_{\text{opt1}})^2}, \quad (\text{C2})$$

$$f_{T2} = \frac{(T - T_{\min2}) \times (T - T_{\max2})}{(T - T_{\min2}) \times (T - T_{\max2}) - (T - T_{\text{opt2}})^2}. \quad (\text{C3})$$

The soil pH factor is calculated as



$$f_{\text{pH}} = \frac{(\text{pH} - \text{pH}_{\min}) \times (\text{pH} - \text{pH}_{\max})}{(\text{pH} - \text{pH}_{\min}) \times (\text{pH} - \text{pH}_{\max}) - (\text{pH} - \text{pH}_{\text{opt}})^2}. \quad (\text{C4})$$

Here,  $T$  is the soil temperature ( $^{\circ}\text{C}$ );  $Q_{10}$  represents the temperature sensitivity in different conditions, including  $\text{DOCprod}Q_{10}$ ,  $\text{Aceprod}Q_{10}$ ,  $\text{CH}_4\text{prod}Q_{10}$ , and  $\text{CH}_4\text{oxid}Q_{10}$ , all of which have a value of 2.5;  $T_{\min 1}$  and  $T_{\min 2}$  are the minimum soil temperature ( $^{\circ}\text{C}$ ), whose values are 0 and 20, respectively;  $T_{\max 1}$  and  $T_{\max 2}$  are the maximum soil temperature ( $^{\circ}\text{C}$ ), whose values are 20 and 50, respectively; and  $T_{\text{opt}1}$  and  $T_{\text{opt}2}$  are the optimum soil temperature ( $^{\circ}\text{C}$ ), whose values are 10 and 35, respectively. pH is the pH value, and  $\text{pH}_{\min}$ ,  $\text{pH}_{\max}$ , and  $\text{pH}_{\text{opt}}$  are the minimum, maximum, and optimum pH, with values of 3, 9, and 6.2, respectively. Soil pH factor was calculated with reference to Cao et al. (1996) and Meng et al. (2012). The soil moisture factor  $f_{\text{moist}}$  was adopted from the IBIS.

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